HANDOUT OF ARBAMINCH UNIVERSITY
OCEANOGRAPHY & MARINE METEOROLOGY

BY

Prof.(Dr.)A.S.N.MURTY

HANDOUT OF ARBAMINCH UNIVERSITY

INSTRUCTOR: Prof.(Dr.)A.S.N.MURTY

OCEANOGRAPHY & MARINE METEOROLOGY
MHD 0904
B.Sc. II Year: II Semester

CHAPTER – I: Ocean Dimensions

1. Comparative dimensions of Earth & Oceans:

Table of Earth Dimensions

Dimension	Magnitude	Units
Mass	6×10^{27}	Kg
Volume	1.1×10^{12}	Km^3
Circumference	40×10^{3}	Km
Poar Radius	6,356	Km
Equitorial Radius	6,378	Km
Total Surface Area	510×10^{6}	Km^2
Land Surface Area	149×10^{6}	Km^2
Ocean Surface Area	361×10^{6}	Km^2
Ocean Volume	$1,370\times10^{6}$	Km^3
Ocean Average depth	3,795	meters

If all the water (1370 million) is spread uniformly over the surface of the earth (510 million) the mean height of water level would be approximately 2.7 km. This height is called the geoid. This corresponds to the pressure of 270 kg wt cm^{-2}, if the atmospheric pressure is approximately 1 kg wt cm^{-2}.This means the mass of the oceans is 270 times that of the atmosphere.

The mean depth of the oceans is the ratio of the volume of the oceans (1370 million) to the area of the ocean basins (361 million), which is equal to 3.8 km. Which means the mean depth of the oceans is 1/1675 times of the mean radius of the earth. This in turn implies Oceans constitute only a very thin layer on the surface of the earth. Thus the earth and the oceans can be viewed as a plum fruit such that the center seed of this plum fruit is the core, the flesh is the mantle and the skin is the crust of the ocean & land.

Looking at the globe the more of the earth's surface is covered by ocean than by land, i.e. Ocean : Land is about 71 : 29. Furthermore the proportion of water to land in the Southern

Hemisphere is more (4:1) than the Northern Hemisphere (1.5:1).In area the Pacific Ocean is as large as the combined area of the Atlantic and the Indian Ocean. While the Pacific Ocean is nearly 46% of the total world oceans, the Atlantic and the Indian Ocean respectively are 23% and 20%. The remaining seas and oceans are 11% of the world oceans.

While only 11% of the land surface is more than 2000 meters above sea level, 84% of the sea bottom is more than 2000 meters deep.

The highest peak on the earth's surface (Mount Everest, Himalayas in Nepal) is only 8840 meters above sea level whereas the deepest portion in the oceans (Mindanao trench in western Pacific in the Mariana region) is 11,524 meters below sea level as shown in Fig.1. This means that oceans are much deeper than the land is high.

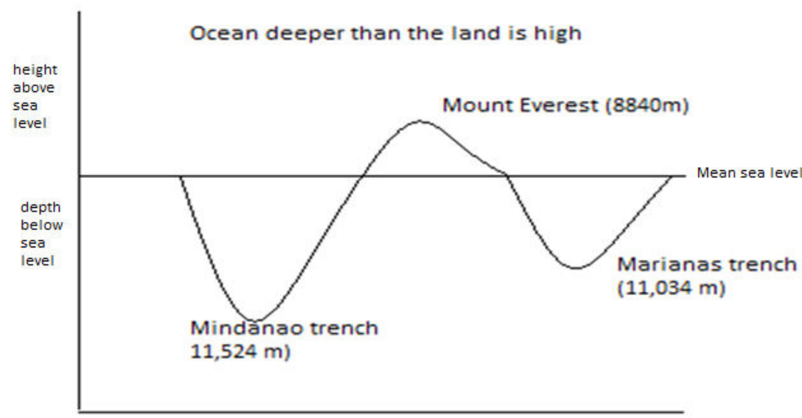


Fig.1. Comparison of land height and ocean depth

2. OCEAN TOPOGRAPHY:

Starting from the land towards the ocean, the main divisions are the beach, the shore, the continental shelf, the continental slope, the continental rise and the abyssal sea bottom (Deep Sea) as shown in the Fig.2.

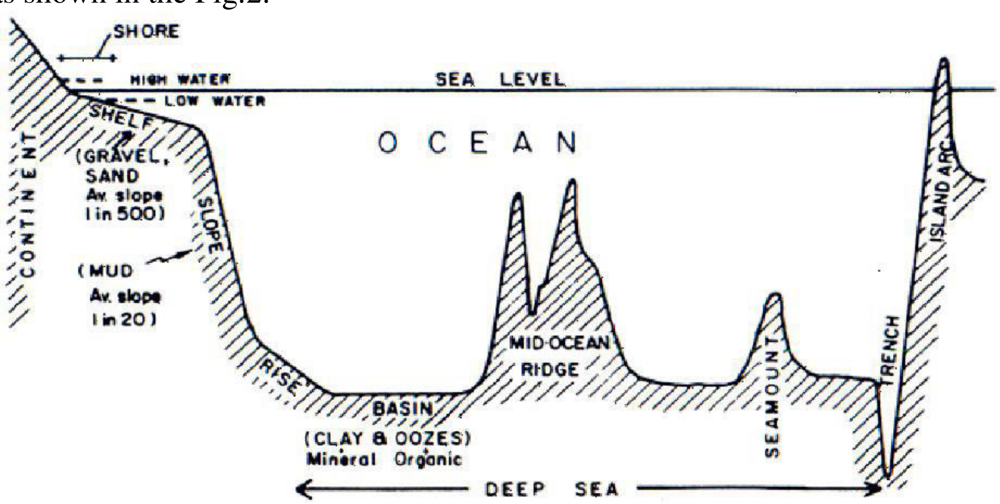


Fig.2. Divisions starting from land to ocean

The shore is defined as that part of the land mass close to the sea which has been continuously modified by the action of the sea due to swash and back wash.

1.1.The Beach:

The main features of the beach are shown in Fig.3 as below. The beach is defined as ‘ the zone of unconsolidated sand material extending landward from the mean low water line to the place of permanent vegetation.

Though the beach, in general, remains in equilibrium condition, it is subjected to short-period changes of erosion or deposition seasonally. But now a days man made constructions like jetties, piers, groins or construction of sea port cause large scale changes and causes distability of the natural environment of the beaches.

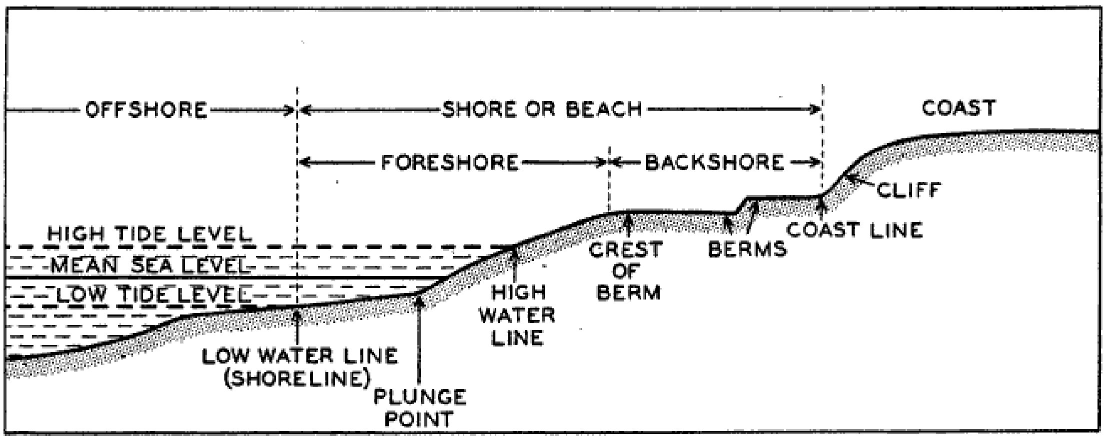


Fig.3. Different parts of the Beach & its terminology

1.1.1. Uses of beaches & continental margin:

Beaches are used today mercilessly for various purposes like tourism, industry, urbanization, recreation, transportation, aquaculture etc. As a result it causes large scale pollution which results in vanishing of the endangered marine animals like horse shoe crab, sea horse, star fishes etc.

The industrial uses of beaches are salt panning, extraction beach placer deposits like rare earth minerals of illiminite, rutile, zircon which are useful for extraction of radioactive substance thorium which is useful for atomic power plants and atomic reactors. Later oil and natural gas both in the fore shore and off shore, sand mining for construction works. Some countries like France and Japan have setup tidal power plants on the beaches. Another important industry is establishment of oil refineries and cargo handling in the sea port areas.

The agricultural uses are aqua culturing of fishes, prawns, crabs, eels etc., growing of social forestry and commercially useful plants like coconut and cashew plantations and mangrove forests in estuarine regions.

Most of the beautiful beaches are converted today as touristic resorts to attract tourists as the local government gets lot of revenue by the tourism industry. This is done by creating marine parks, boating, surfing and multi star hotel complexes.

The continental shelf, slope and rise together is called continental margin. It is within this continental Margin, that virtually the entire ocean’s living and nonliving resources are lying. While the important non living organic resources like oil and natural gas are in the Continental Margin area, the

manganese nodules containing nickel, copper and cobalt are in the deep sea bed. Apart from these resources, large placer deposits of 'rare earth minerals' like zircon, illimanite and Rutile are present on the beaches.

1.2. The continental shelf:

It extends seaward from the shore with an average gradient of 1/500 as shown in Fig.2. At its bottom, the gradient increases to about 1/20 near the continental slope. The shelf has an average width of 65 km. But in some places it is lesser wide and in other places of the oceans it is more wide. The average depth is about 130 m. Most of the world fisheries are located on the continental shelf.

1.3. The continental slope:

It goes down from the end of shelf to about 4000 m or so till the deep ocean bed touches. The slope is often as narrow as 20-30 km in width. The Continental Rise (Fig.2&4), a part of the slope, is an area largely composed of continental sediment that has been transported down the slope through the ages and which merges gradually with the true ocean bottom. The shelf, slope and the rise will make-up the so called Continental Margin. It is within the continental margin, as thus mentioned, that virtually all of the oceans' non living organic and resources (oil and natural gas) are thought to lie. Very typical features of the shelf and the slope are the sub marine canyons which are of worldwide occurrence. They are 'V' shaped valleys or deep gorges and are usually found off the coast with river mouths and so are thought to have been formed due to river cuts.

1.4. The abyssal plain (the deep sea bottom):

It is the last and most extensive area with highly variable topography. Depths of 3000 to 6000 m are found over 76% of the deep ocean basins. The deep sea bottom is not flat and contains mountains, valleys, plains just as on land as shown in Fig.4. The Mid Oceanic Ridge system is probably the most extensive single feature of the Ocean's topography. Starting south of Greenland it extends along the middle of the Atlantic from north to south spreads through whole of the Indian and the Pacific Oceans. On the abyssal plains in some areas of the oceans extensive deposits of small burnt-baked-potato-like objects called Ferro-manganese nodules have been recently discovered which are rich in manganese and iron. It appears that if they are extracted the world's needs of iron and manganese can be solved. The economic viability of extraction of these nodules is being seriously researched all over the world now.

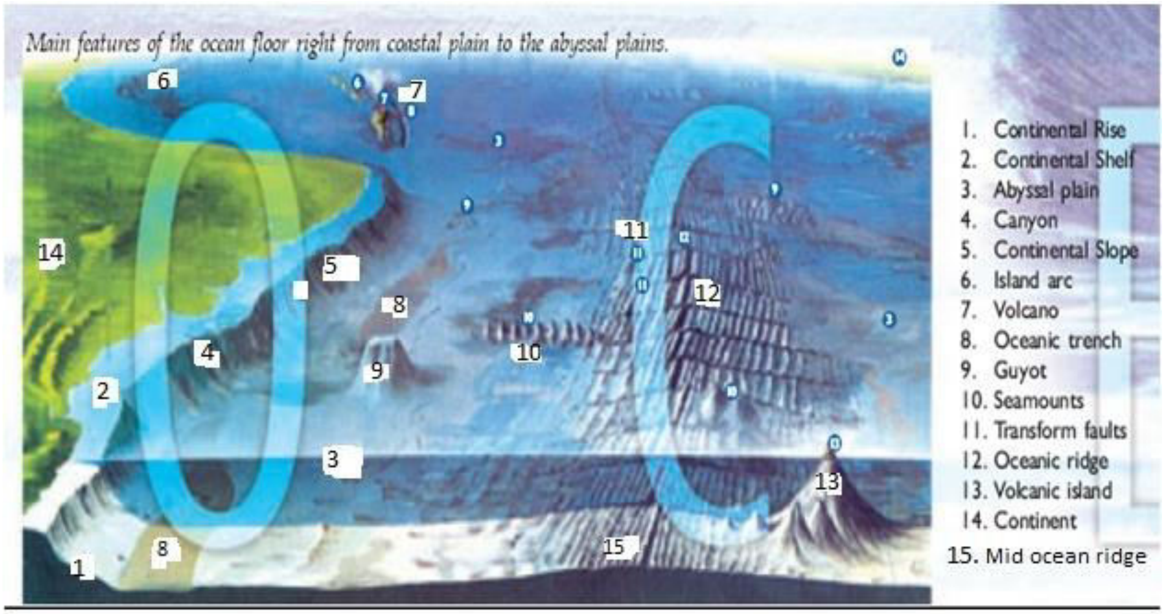


Fig.4: The main features of ocean floor.

HANDOUT OF ARBAMINCH UNIVERSITY

INSTRUCTOR: Prof.(Dr.)A.S.N.MURTY

OCEANOGRAPHY & MARINE METEOROLOGY
MHD 0904
B.Sc. II Year: II Semester

Chapter II

PHYSICAL PROPERTIES OF SEA WATER:

Water is one of the most remarkable compounds in nature. Most of the processes of man kind's environment are ultimately depend on its unique physical properties. The liquid state of water is very rare in the universe.

2.1.0. BASIC PURE WATER CHARACTERISTICS:

Compared to other chemically related compounds water behaves physically in a unique manner. Water has anomalous behavior. All material substances expand when heated and contract when cooled. But water follows this rule only partly. At temperatures below 4^oC it expands with further cooling it freezes at 0^oC. If this abnormality were not existed, ice would have sunk down to the bottom instead of floating at the surface which acts as a protective shield to prevent further freezing. If the ice were to sink, oceans would not have existed in liquid form, as all the polar waters would have frozen within no time and the water goes to those regions will not return again to lower latitudes.

2.1.1. What is the reason for this anomalous behavior?

If water were a normal compound like Ammonia, Hydrogen Fluoride, HCl etc., its theoretical boiling point would have been around -80^oC and freezing point around -150^oC. All this peculiar behavior of water is due to its molecular structure. Since the water molecule consists of two atoms of hydrogen and one atom of oxygen which are not attached on either side of it at 180^0 as shown in Fig.1

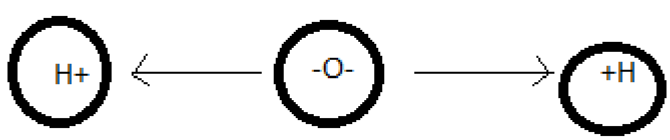


Fig.1.This is not the actual structure of water molecule

But the two valences of oxygen atom join the values of two hydrogen atoms at an angle of 105^0 (as shown in Fig.2) is the real structure. The two H^+ atoms are positively charged, the $^-O^-$ is double negatively charged and the above asymmetric structure (Fig.2) with the displacement of electric charges results in the formation of a strong dipole moment.

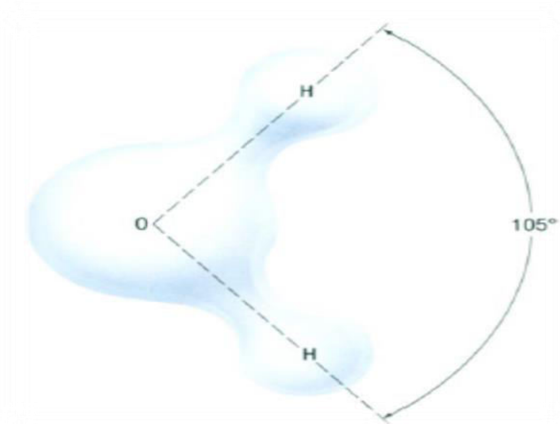


Fig.2.Real structure of water molecule

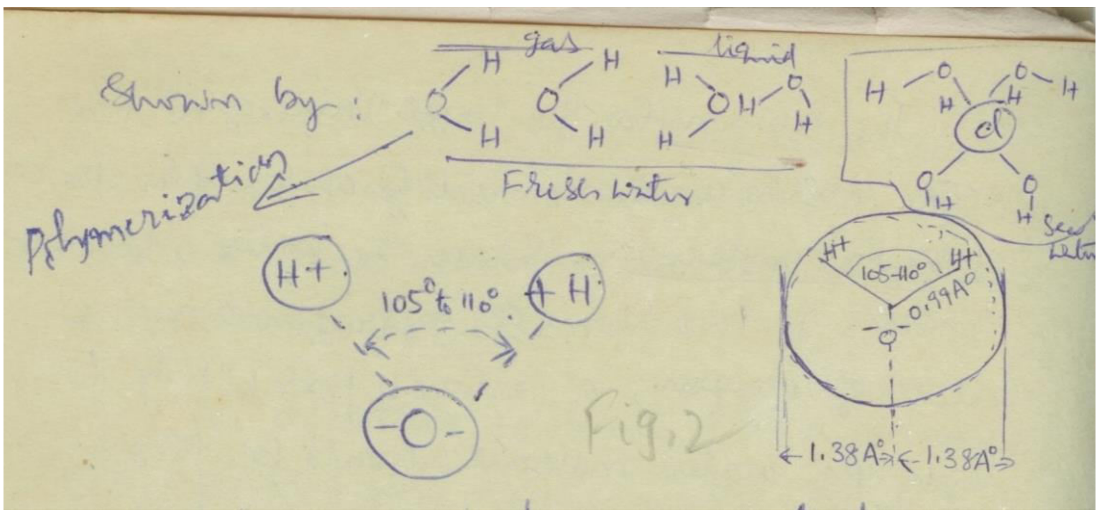


Fig.3. Polymerization and angular structure of water molecule

Sea water acts as a condenser with pure water as dielectric as Na^{+}- H_2O – Cl^{-}, where Na^{+} is cation and Cl^{-} is anion as shown in Fig.4.

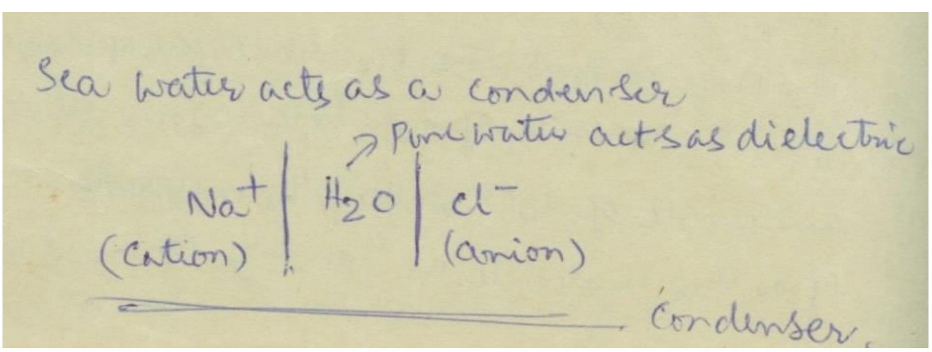


Fig.4. Sea water as condenser

Because of the highest dielectric constant (81), water has another important property, that is, it has the highest dissolving power that is the reason why it dissolves many substances. The strong dipole moment is the cause for the existence of strong forces interacting between the molecules themselves. This leads to the formation of molecular groups called ‘polymerization’(Fig.3). These types of groups give rise to three different water structures. They are tetra hedral structure, Quartz like lattice structure and Ball pack structure.

While the tetrahedral structure occupies the least density, ball pack structure occupies the highest density. The tetra hedral structure is shown in Fig.5.

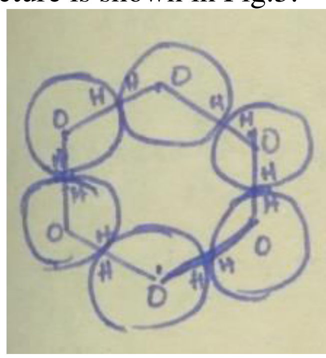


Fig.5.Tetrahedral structure

2.2. SALINITY.

Did you ever wonder why the oceans are filled with salt water instead of fresh? Just where did the salt come from? Most of the salt in the oceans came from land. Over millions of years, rain, rivers, and streams have cut and washed the rocks containing different salts and carried them into the sea. Some of the salts in the oceans came from undersea volcanoes and hydrothermal vents. When water **evaporates** from the surface of the ocean, the salt is left behind. After millions of years, the oceans have developed a noticeably salty taste.

Different bodies of water have different amounts of salt mixed in. That is why different oceans have different salinities. Salinity is expressed by the amount of salt found in 1,000 grams of water. Therefore, it is denoted as one part per mille (‰).

While the salinity of all oceans of the globe varies between 32to 37‰, the average ocean’s surface salinity is 35‰.

Sea water is a complicated solution and contains the majority of the known elements. Among the 100 parts of sea water 96.5 parts consist of water and the rest 3.5 parts consist of dissolved materials which are mostly different salts.

Out of the total dissolved material the distribution of different ions are: Chloride ion is 55%, Sodium is 30.6%, Sulphate is 7.7%, Magnesium is 3.7% and Potassium is 1.1%. This total amount of dissolved material in sea water is termed as ‘salinity’.

Salinity is defined as ‘ the total amount of solid material in grams contained in one kilogram of sea water when all the carbonate has been converted to oxide, the bromine and iodine are replaced by equivalent amount of chlorine and all the organic matter is completely oxidized’.

If the average salinity of the ocean is taken as 35‰, the total amount of salt in the ocean works out to 4.84×10^{16} tons. This corresponds to a volume of 21.8 million km^3 which when spread evenly over the sea (361×10^6 km^2) would be a layer of salt of 60 m thick.

A significant feature of sea water is that while the total concentration of dissolved salts varies from place to place, the ratios of the more abundant components remain almost constant. This is called constancy of composition.

2.2.1: DETERMINATION OF SALINITY:

The most accurate and easiest way of estimation of salinity is by chemical method. That is by titration of sea water with $AgNO_3$ using potassium chromate ($K_2Cr_2O_3$) or potassium dichromate (K_2CrO_4) as indicator. The reaction is

$$NaCl + AgNO_3 \rightarrow AgCl + NaNO_3$$

Which means only the amount of chlorine or chlorinity of sea water is estimated with the help of Silver Chloride (AgCl) precipitate. So salinity can be estimated from chlorinity using the relation:

$$S‰ = 1.80655 \times Cl$$

Because of this situation, chlorinity is defined as 'the total amount of chlorine in grams present in one kilogram of sea water when bromine and iodine are replaced by an equivalent amount of chlorine.

Salinity can also be estimated using refractive index and conductivity of sea water. The salinometer uses the principle of conductivity.

2.3.0. DENSITY OF SEA WATER:

The density of a substance is defined as the mass per unit volume. So the unit in C.G.S system is gram per c.c (gm cm^{-3}). The specific gravity is defined as the ratio of density to that of the distilled water.

The density of sea water depends on three variables temperature(t), salinity (S), and pressure (P) at which it is collected and is called the density '*in situ*' which is denoted as $\rho_{s,t,p}$.

The values of open ocean density vary from about 1.02400 to 1.03000 gm cm^{-3}. As a matter of fact the actual variation of density of sea water is very less i.e. it varies from the second decimal place onwards and this small variation is very important. So to represent only the last four decimal places, it is converted using the relation

$$\sigma_{s,t,p} = (\rho_{s,t,p} - 1)\, 10^3 \quad \ldots\ldots\ldots\ldots\ldots\ldots(1)$$

Where $\sigma_{s,t,p}$ is called specific density *in situ*.

As pressure in the ocean is related to the depth and is converted using 1 db = 1 meter when a sample of sea water is collected at different depths, pressure effect is automatically included in it. So the quantity 'σ' at a particular depth can be referred only in terms of salinity and temperature and at the sea surface it is $\sigma_{s,t,0}$ which is called as 'σ_t'. So σ_t is a function of temperature and salinity only.

The relationship between σ_t and salinity and temperature is a complicated nonlinear function. So yet another parameter called σ_o has been defined which is a function of salinity only (at t = 0 and p = 0) and is directly related to salinity as

$$\sigma_0 = a + b S + c S^2 + d S^3 \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots(2)$$

Where a = -0.093, b = 0.8149, c = -0.000482, d = 0.0000068.
Or in terms of chlorinity it can be written as:

$$\sigma_o = -0.069 + 1.4708\, Cl + -0.001570\, Cl^2 + 0.0000398\, Cl^3.$$

The values of σ_o, Cl and S‰ are computed using these formulae and given in Knudsen's Hydrographic tables.

Further σ_t and σ_o are related as : $\sigma_t = \sigma_o - D$.

Here D is expressed as a complicated function of σ_o and thermal expansion and compressibility of sea water.

2.3.1. SPECIFIC VOLUME:

The specific volume $\alpha_{s,t,p}$ is the reciprocal of density in *situ* $\rho_{s,t,p}$. This means

$$\alpha_{s,t,p} = \frac{1}{\rho_{s,t,p}}$$

Generally in oceanography to avoid large number of decimals, the specific volume is expressed as an anomaly 'δ' as

$$\delta = \alpha_{s,t,p} - \alpha_{35,0,p}$$

where $\alpha_{s,t,p}$ is specific volume in *situ* and $\alpha_{35,0,p}$ is the specific volume of an ideal standard ocean which has salinity 35‰ and temperature 0^oC.

Although specific volume anomaly has been defined, still this parameter is also a complicated function of t,s and p. To avoid this complicacy, it is further broken down as

$$\delta_{s,t,p} = \delta_s + \delta_t + \delta_{s,t} + \delta_{s,p} + \delta_{t,p} = \Delta_{s,t} + \delta_{s,p} + \delta_{t,p} \quad \ldots\ldots\ldots(3)$$

Thermosteric anomaly:

The thermosteric anomaly is defined as the anomaly of specific volume that would be attained if the water were changed isothermally to a standard pressure of one atmosphere.

In other words the sum of the first three terms ($\Delta_{s,t}$) on the right hand side of equation(2) is termed as thermosteric anomaly and said that to study the upper layers of the oceans, this parameter is sufficient.

$$\Delta_{s,t} = \delta_s + \delta_t + \delta_{s,t}$$

The thermosteric anomaly is related to σ_t as $\Delta_{s,t} = 0.02736 - \dfrac{10^{-3}\sigma_t}{1+10^{-3}\sigma_t}$ ……..(4)

2.4. THERMAL PROPERTIES:
ADIABATIC TEMPERATURE CHANGES OF SEA WATER:

The principle of adiabatic gain and loss of heat on compression and expansion of gases provide the basis for refrigeration and air conditioning.

If a work is done on the parcel of sea water without the gain or loss of heat to its environment is called adiabatic process. This brings the concept of potential temperature (θ).

Potential temperature is defined as the temperature which the fluid would attain if brought adiabatically from its actual depth to the sea surface. It is thus different from *in situ* temperature which is the temperature of the parcel measured at its actual depth.

If a water parcel is raised from a deeper layer to surface, the sea pressure decreases, and so the volume of water sample expands. As a result the temperature of the parcel decreases. This effect is called adiabatic cooling and the associated temperature is called potential temperature (θ). Conversely if a parcel goes from surface to deeper layers, adiabatic heating takes place.

2.5. COLLIGATIVE PROPERTIES:

The colligative properties are the unique properties of solutions. Sea water is also a solution because it contains dissolved salts and water.

The magnitude of colligative properties depends on the concentration of ions in the solution and their activity. The colligative properties are:
1. Elevation of boiling point, 2.Elevation of osmotic pressure, 3. Elevation of surface tension, 4.Lowering of freezing point, 5. Lowering of vapor pressure and 6. Lowering of temperature of maximum density.

2.5.1. Lowering of temperature of maximum density:

Pure water has a maximum density at 4^{o}C. With increasing salinity the temperature of maximum density of sea water decreases as shown in Fig.6. And with increasing salinity the temperature of freezing point also decreases. Both the curves meet at a point (24.67,-1.33).

Thus the rate of decrease of maximum density is greater than that of freezing point upto this point (24.67,-1.33) beyond that it decreases quicker than the other one.

But at a salinity of 24.67‰ the temperature of maximum density and temperature of freezing point both are same (-1.33^{o}C).

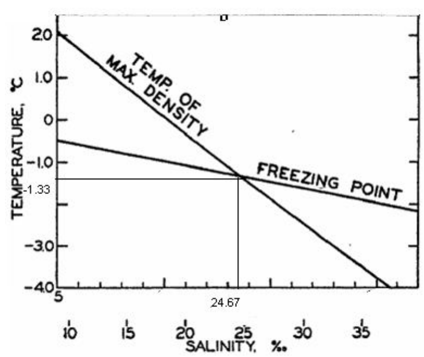


Fig.6. The variation of temperature of max density and freezing point with increase of salinity

2.6.0. OPTICAL PROPERTIES OF SEA WATER.

2.6.1. Extinction of radiation:

The radiation that penetrates into the surface layers of the ocean depends upon absorption of pure water, suspended and dissolved substances, scattering by suspended and dissolved materials.

Two important terms, radiance and irradiance are usually used in this regard.

Radiance: Energy per second received by the sea surface from a certain direction.

Irradiance: Energy per second received by the sea surface from all directions that comes usually in the horizontal.

The Fig.7 shows the penetration of solar energy and absorption of different wavelengths in the ocean.

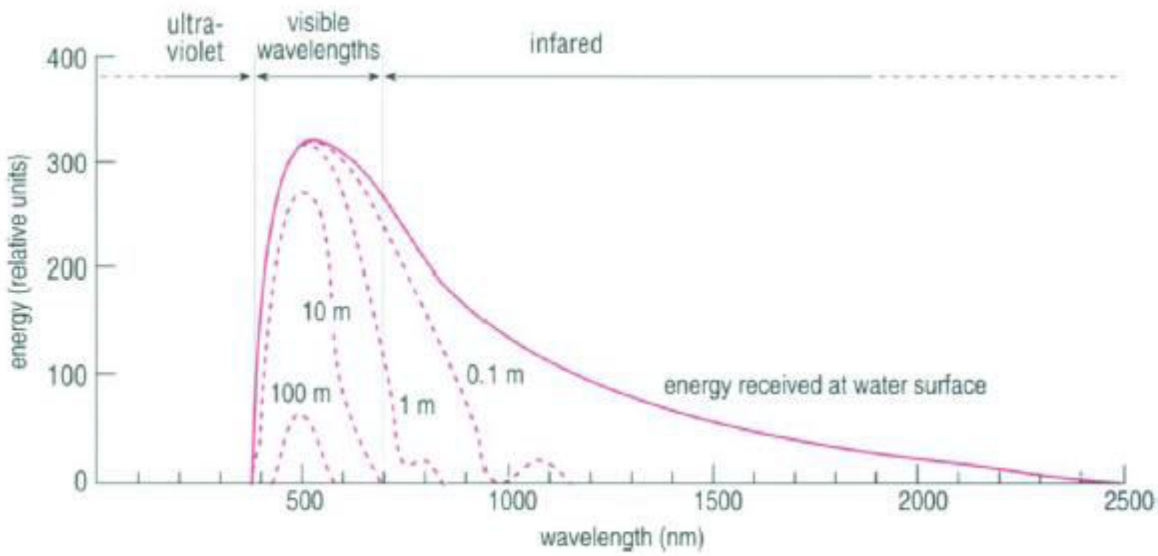


Fig.7. Penetration of solar energy at different depth in the ocean (0.5μ= 500 n.m; 1 μ = 10^8 n.m)

Fig.7 shows the shorter wave lengths (i.e. those near the blue end of the visible spectrum) penetrate deeper than longer wave lengths of red etc. So the color of light below the sea surface is blue-green. Infrared radiation is the first to be absorbed, followed by red and so on. The total energy received at a given depth is equal to the area enclosed by the curve of that depth. The 100 meter curve occupies only 1/50 th of the surface curve. All of the infrared radiation is absorbed within about a meter of the surface, and nearly half of the total incident solar energy is absorbed with in 10 cm of the surface.

Penetration of light in waters will also depend up on the clarity or transparency of the water, which in turn depends on the amount of suspended matter in it.

2.6.2. EXTINCTION COEFFICIENT:

The diminishing effect of radiation while penetrating into different layers of ocean is known as extinction of radiation. The extinction is due to the attenuation of radiation which is the total effect of absorption, scattering and reflection of radiation penetrating into the layer.

In oceanography the greater interest is attached to the rate at which the downward traveling light decreases. This rate of decrease can be defined by a coefficient called extinction coefficient ‘K’ and is given by the equation:

$$I_z = I_0 e^{-kz}$$

Where I_o is the amount of sun light intensity reaching the sea surface. I_z is the light intensity reaching at z meters and 'k' is the extinction coefficient.

$$-Kz = \ln(I_z) - \ln(I_o)$$

$$\text{or} \quad K = (1/Z)\, 2.303\, (\log I_o - \log I_z)$$

If Z = 1 meter, $K = 2.303 (\log I_o - \log I_z)$.

The factor 2.303 is due to change of natural logs to base 10 logs. Z is generally taken as one meter because it is easy for calculation. The extinction coefficient can be measured with a Sacchi disc and hydrophotometer.

2.6.3. VARIATION OF 'K' IN DIFFERENT WATERS:

The variation of extinction coefficient of solar radiation with wave length in pure water and in different types of water is shown in Fig.8.

The selective extinction of day light in the whole range of the visible part of the spectrum in Fig.8 shows remarkable differences with different water types. With pure water the extinction is the strongest in the range of longer wave length.

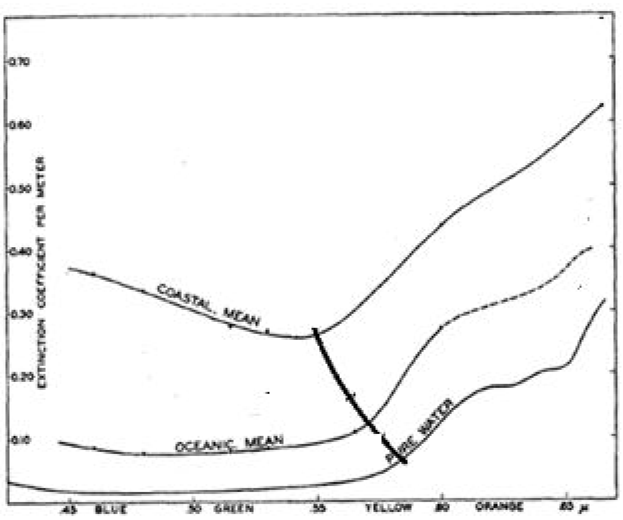


Fig.8. variation of extinction coefficient in different waters. Note the curved line of shifting to longer wave lengths.

With increasing turbidity of the open sea water, and especially of coastal water, the extinction in the short wave length (blue) (0.47- 0.48 μ) becomes more significant. The selective extinction produces a shift of the minimum extinction coefficient from shorter to longer wave lengths as the turbidity increases. Clear open ocean water is most transparent for blue light of about 0.47 to 0.48 μ whereas in turbid coastal water the maximum transparency has shifted to about 0.55μ (the yellow-green). Note the shifting of the minimum extinction coefficient to longer wave lengths as it goes from turbid coastal waters to clear oceanic waters as shown by the thick vertical curved line in Fig.8.

That is why the color of the oceans looks blue. While the largest part of the surface of the ocean in the tropics and sub tropics is blue, the coastal areas, shallow seas and Polar Regions are greenish in color. If there is green phytoplankton their chlorophyll content will absorb the blue light and shift the water color to green. In coastal regions and river mouth areas, rivers bring large sediment containing organic substances, so coastal waters appear brown or yellowish green color.

2.7. ACOUSTICAL PROPERTIES OF SEA WATER

The most important acoustical parameter of the ocean is the sound velocity. The transmission of sound with low attenuation in the ocean enables scientists to use acoustical methods to determine the depth of the oceans, sea bottom structure and to locate the hidden submarines as well as to communicate over considerable distances in a medium that is nearly opaque to most electromagnetic radiation. That is how all the marine animals can use sound as a communicating device.

2.7.1. SOUND PROPAGATION IN THE VERTICAL:

The sound velocity can be used to determine the depth of a shoal, lurking enemy submarine or a school of fish. As there are different regions in the oceans, sound travels at different speeds in different regions as shown in Fig 9. Normally the sound velocity increases with depth in the oceans except in the thermocline region.

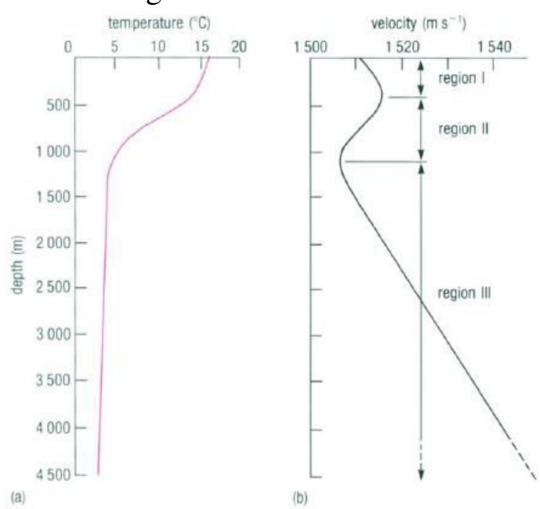


Fig.9. Vertical variation of temperature (a) and sound velocity (b) in the oceans

The ocean contains three layers, namely, mixed layer (region I), thermocline (region II) and stratified layer (region III) as shown in Fig.9 above. In the mixed layer temperature is uniform and pressure increases with depth and so sound velocity increases. In the thermocline temperature sharply decreases with depth and so its influence dominates the increase of pressure and so sound velocity decreases in thermocline (region II). In the stratified layer (region III) again pressure effect dominates as temperature and salinity influence is negligible and so sound velocity increases again.

Because of the property of sound refraction in different regions of the Ocean, a shadow zone in the surface layer and a sound channel in the deep layer are formed as shown in Fig.10. If

the sound source is kept between mixed layer and thermocline, a shadow zone is formed and in the region of minimum sound velocity sound channel is formed which is in region III. This sound channel area is called SOFAR (Sound Fixing And Ranging) channel. The sound pulse sent in this channel will travel several thousands of kilometers without much loss. SOFAR channel can be used by ships in distress. This can be used to measure the deep water currents by using drifting buoys in this channel.

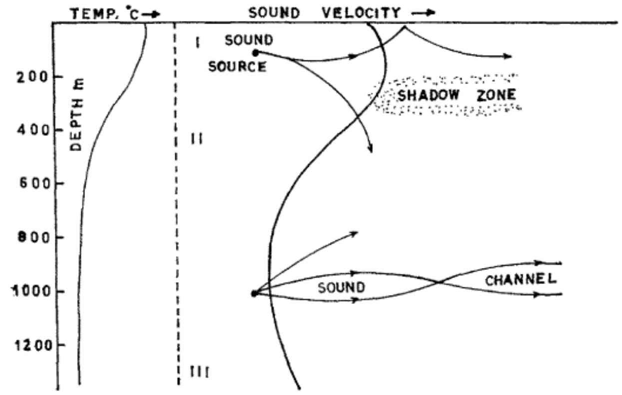


Fig.10. Formation of shadow zone and SOFAR channel in the oceans

2.7.2. DEPTH DETERMINATION USING SOUND VELOCITY:

Because of these differing velocities in different zones, a weighted average of sound velocity is used to determine the depth from the equation:

$D = (½)\ t.v$

where 't' is the time taken for the sound pulse to travel from the ship to the sea bottom and back to the ship as shown in Fig.11, which means it travelled twice the distance.

The average speed of sound used in sonic sounding is about 1500 m/s.

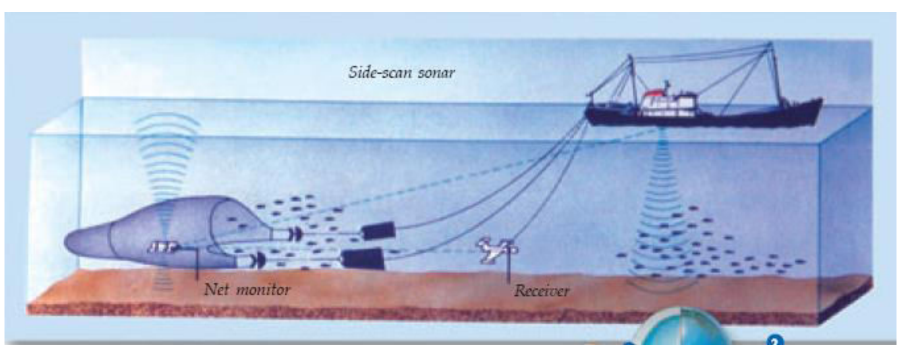


Fig.11.Mapping the sea floor or identifying school of fish using Echo sounder or side scan sonar. Note the distance is twice as the sound pulse travels to the target and come back to the ship.

HANDOUT OF ARBAMINCH UNIVERSITY

INSTRUCTOR: Prof.(Dr.)A.S.N.MURTY

OCEANOGRAPHY & MARINE METEOROLOGY
MHD 0904
B.Sc. II Year: II Semester

CHAPTER – III

DISTRIBUTION OF PHYSICAL PROPERTIES IN THE OCEANS

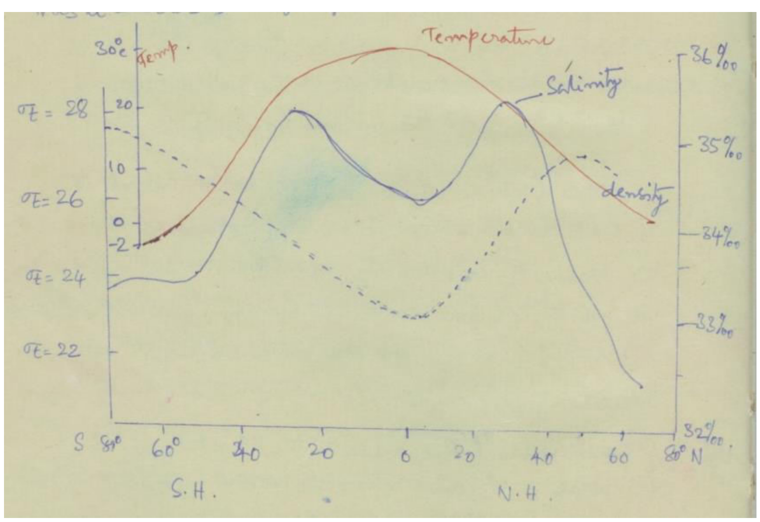
Temperature, salinity, density and oxygen are the three main entities which determine the water characteristics of the oceans. These quantities vary from place to place and time to time in the oceans and from their distribution we can learn about the average circulation of the oceans.

A salient feature of the distribution of many water characteristics is that they are horizontally stratified. In other words, the sea is made up of number of layers as far as these characteristics are concerned, and horizontal changes are much smaller than vertical changes in the same distance.

One feature to be noted regarding the spatial distribution of water characteristics is the value of the property is almost same across the ocean in the east-west direction (zonal direction) but may change rapidly in the north south direction (meridional direction).

3.1. Latitudinal variation of temp, salinity and density:

The horizontal distribution of density in the oceans can be considered as zonal and meridional distribution. While the zonal distribution (east-west) is almost negligibly small, meridional distribution is more. Fig.1 shows the latitudinal variation of temperature, salinity and density (σ_t) at the sea surface. σ_t increases from about 22 near the equator to 26 to 27 at 50^o to 60^o latitude.



.

Fig.1. Latitudinal variation of temperatura, salinity and density in the oceans

Temperature is maximum (28^{0}C) in the equatorial region and least (-2^{0}C) in the polar regions.

Salinity is maximum (37‰) in the sub tropical zones and in the equatorial zone (34.5‰) though high it is less than sub tropics and in the polar zones it is least (less than 33‰). Further the salinity in polar zones of Northern Hemisphere (N.H) (32.5‰) is less than that of Southern Hemisphere (S.H) pole (33.5‰).

With regard to density, it is least (23) in the equatorial zone and highest in the polar zones. The density in the S.H is more than the density of N.H.

3.1.1. Density distribution: Vertical distribution:

With regard to the vertical distribution of density in the oceans, ocean can be considered into three layers. In the equatorial and tropical regions there is usually a shallow surface zone of nearly uniform density called 'mixed layer', then a zone of maximum increase of density called 'pycnocline' and a deep zone of constant high density called 'stratified layer' as shown in Fig.2.

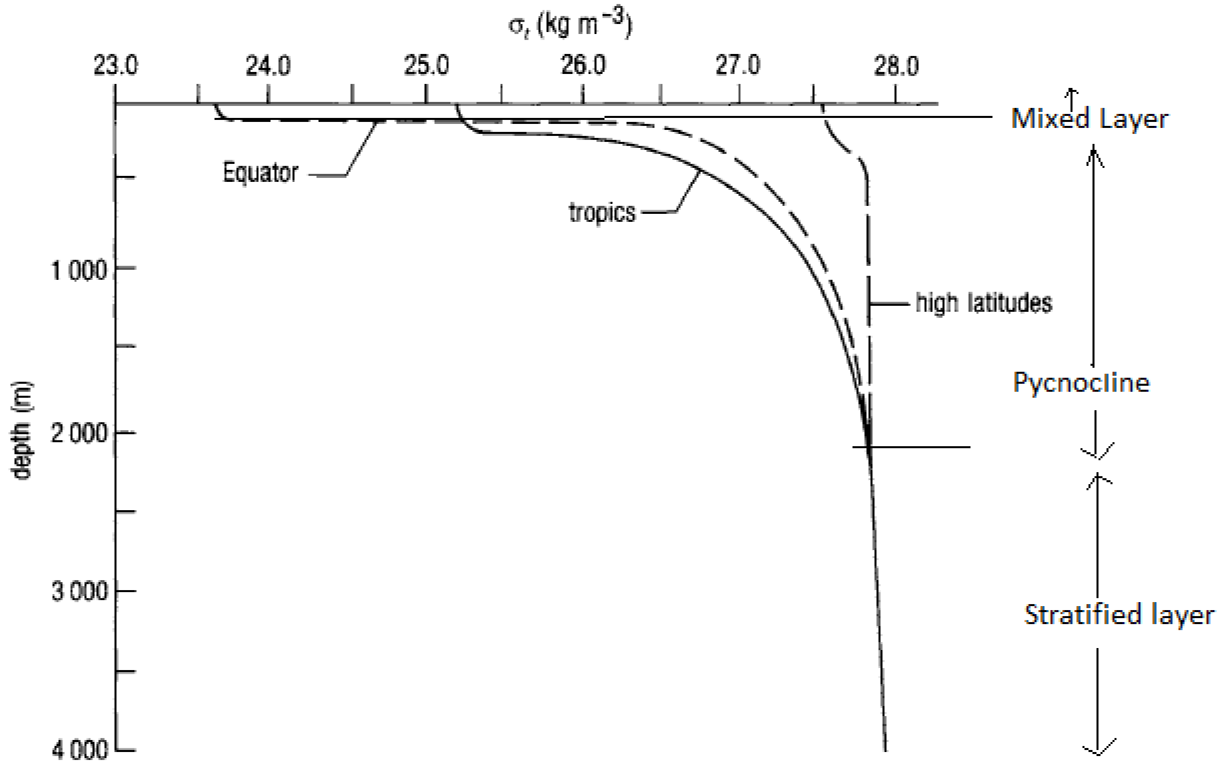


Fig.2. Vertical variation of density in the oceans

However this rate of variation varies considerably between equatorial, tropical and high polar latitudes as shown in Fig.2.

The rate of change of the density with depth ($d\sigma_\theta$ /dz) determines the water's stability.

$$\frac{d\sigma_{\theta}}{dt} \;\lesseqgtr\!=\; 0 \;\; {}^{unstable}_{\;\;stable}neutral$$

If $d\sigma_\theta$ /dz is less than, equal to or greater than zero it is unstable, neutral or stable respectively. Where the stability is high, vertical movement and vertical mixing are minimum, where there is no change of potential density with depth the water is neutrally stable and where it can mix up easily is highly unstable.

3.2. DISTRIBUTION OF TEMPERATURE:

3.2.1. HORIZONTAL DISTRIBUTION OF TEMPERATURE:

With respect to the horizontal distribution of temperature like that of density, the zonal variations are much lesser than that of meridional variations. Though the isotherms, in general, run parallel to the latitudes (Fig.3), very near the continental boundaries, some sharp kinks are observed. Note the sharp bends (kinks) at the coastal boundaries in Fig.3. These sharp kinks are due to land and sea variations (climate variations) and in some places due to advection of cold or warm currents. Also these sharp variations are caused due to upwelling in the eastern shores of the oceans.

The meridional variations of temperature are due to differential heating between equator to pole. Generally the maximum temperature occurs in the equatorial and tropical regions and minimum temperature occurs at the poles as shown in Fig.1.

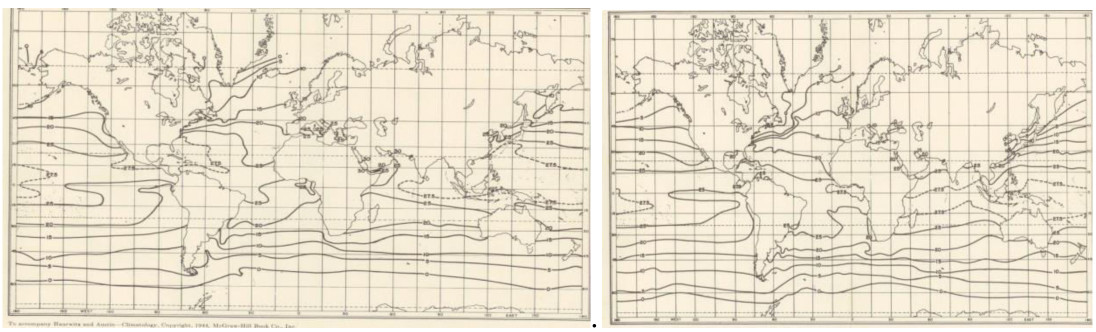


Fig.3. Horizontal distribution of temperature in the oceans in August and February respectively from left to right.

3.2.3. Vertical distribution of temperature:

The ocean can be divided into three vertical zones, basing on temperature in different latitudes as shown in Fig.4.

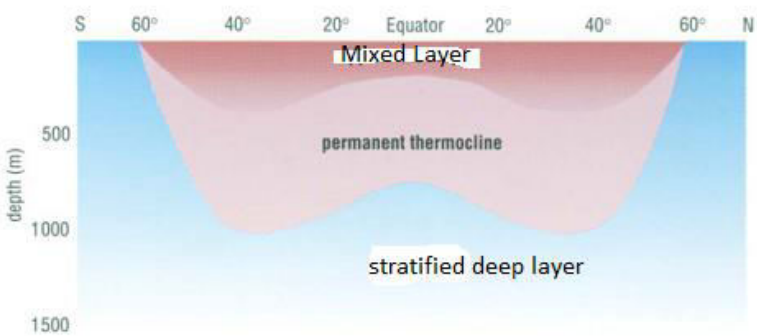


Fig. 4.Vertical layers in the ocean according to temperature

The top layer is the **Isothermal layer**, or mixed layer. This layer is the most easily influenced by solar energy, wind and rain and so maintains higher temperature and uniform in the top 50 meters or so due to mixing. The next layer is the **thermocline**. The thermocline is a zone of sudden and sharp decrease of temperature with increase of depth. The thermocline is generally observed in low and middle latitudes and in polar High latitudes thermocline disappears and in its place dicothermal layer appears as shown in Fig. 5 (High latitudes). Generally, a seasonal upper thermocline above the permanent thermocline develops in middle latitudes. In high polar latitudes the dicothermal layer develops in place of thermocline because a cold water is sandwiched in between two warm waters above and below due to prevention of cooling in the surface waters by presence of ice at the surface. Then the third layer is the **stratified deep water layer**. Water temperature in this zone decreases slowly as depth increases. Water temperature in the bottom parts of the ocean is about 2°C.

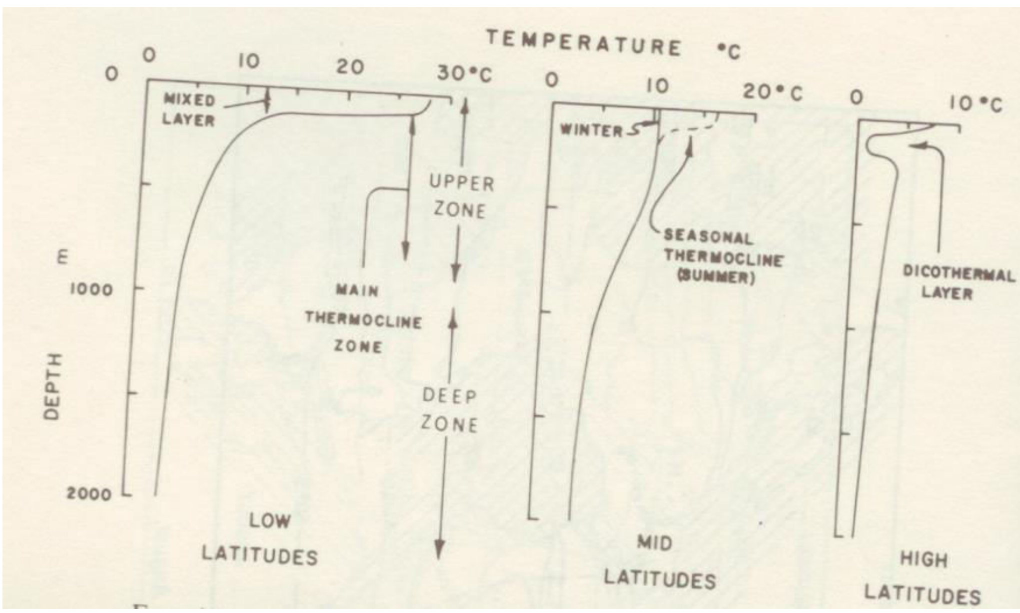


Fig. 5. Vertical distribution of temperature in different latitudes

3.2.4. DISTRIBUTION OF SALINITY:

The total salt content presently dissolved in the ocean is about 5×10^{19}Kgs and every year nearly 2.73×10^{12} Kgs of salt is added to the sea through rivers from land and from mantle and crust of ocean bottom.

Although the average salinity is put at 35‰, numerous salinity differences are encountered in horizontal and vertical directions. The factors that influence the salinity in the oceans are as below:

Increase	Decrease
1. Evaporation	Precipitation
2. Ice formation	Ice melting
3. Advection of high saline water through high saline currents	Advection of low saline water through low saline currents.
4. Mixing with more saline deep water (turbulence & convection)	Mixing with less saline deep water
5. Dissolving of new salts due to deep sea oozes and sub marine volcanic eruptions.	In flow of fresh water due to rivers, glaciers and icebergs.

The average surface salinity distribution over the global oceans with latitude (solid curve marked ‘S’) is shown in Fig.6. It has a minimum just north of the equator and maxima are found in the subtropics (30^{o} N & S) and later decreases to high latitudes. Observations make it clear that surface salinity variation depends on (E-P) factor. Here ‘E’ is evaporation and ‘P’ precipitation. If (E-P) is positive in an area, salinity will increase and if (E-P) is negative, it decreases salinity. The salinity maxima in subtropics may be because evaporation exceeds precipitation (dashed curve), while the equatorial minima are due to precipitation increases evaporation. Note when E-P>0, salinity is maximum and when E-P<0, salinity is minimum in the figure 6.

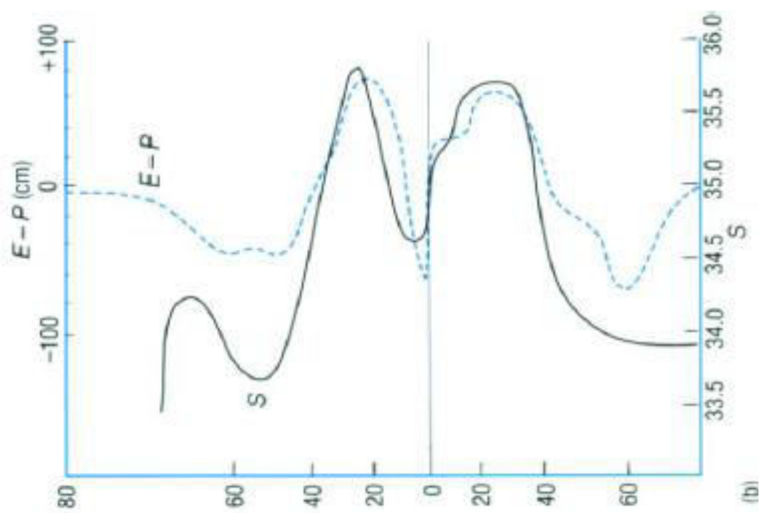


Fig.6. Distribution of surface salinity (solid curve ‘S’) and Evaporation minus precipitation (E-P) (dashed lean curve) in the oceans.

The surface salinity values in the open ocean vary from 33 to 37‰ as shown in Fig.7. Lower values occur locally near the coasts where large rivers empty and in the Polar Regions where ice melts. Higher values occur in the regions of high evaporation such as the eastern Mediterranean Sea (39‰) and the Red Sea (41‰) and the Persian Gulf (40‰). On an average the North Atlantic is the most saline at the surface (35.5‰), the South Atlantic and the South Pacific less (35.2‰) and the North Pacific the least saline (34.2‰) among the major oceans.

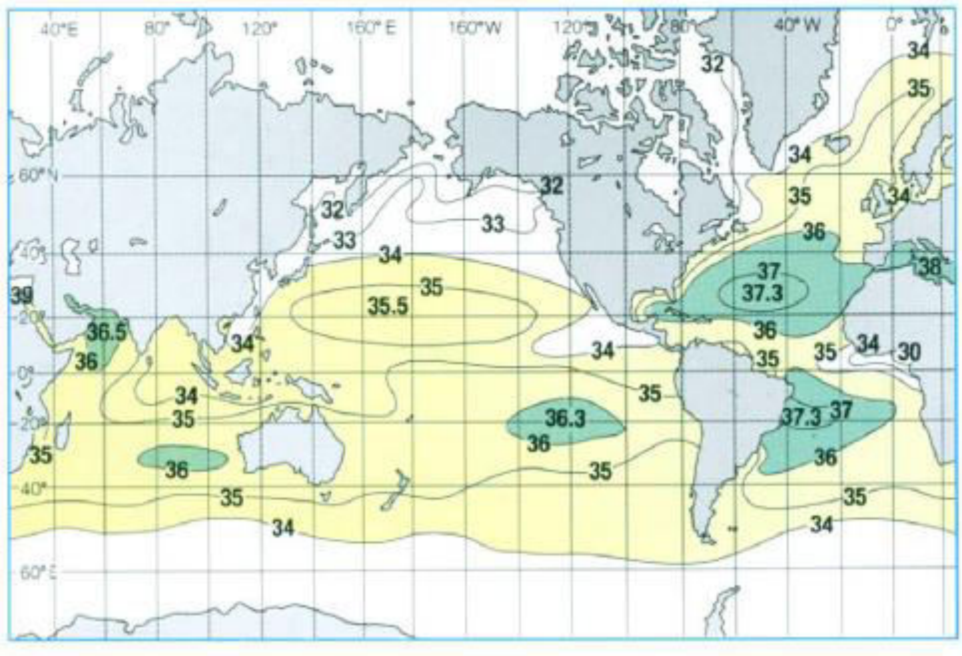


Fig.7. Distribution of surface salinity in the oceans

HANDOUT OF ARBAMINCH UNIVERSITY

INSTRUCTOR: Prof.(Dr.)A.S.N.MURTY

OCEANOGRAPHY & MARINE METEOROLOGY
MHD 0904
B.Sc. II Year: II Semester

CHAPTER-1V
HEAT BUDGET OF THE OCEANS

The main source of energy for the earth and its environment is the sun. At the outer limit of the earth's atmosphere the intensity of solar energy incident on a surface at right angles to the solar beam is nearly equal to 2 gm.cal.cm^{-2} min^{-1}(langleys per day). This is called solar constant. The earth intercepts a circular sample of this radiation which has an area of πR^2. The earth's rotation distributes this sample of energy over the spherical surface of the earth, which has an area of $4\pi R^2$. As a result the average input of solar radiation at the top of the atmosphere is $\pi R^2/4\pi R^2$ = 0.5 gm.cal.cm^{-2} min^{-1} (ly/day).

Our experience shows that in general the average temperature of the earth and oceans is maintained constant which amounts that the gains and losses of heat by the oceans are balanced with each other. This balance of losses and gains of heat is called the heat budget.

The doubt here is while the low latitudes receive surplus radiation and high latitudes receive deficit radiation on a daily basis how the mean temperature of the earth and oceans are maintained constant over a year. Therefore there must be transport of heat from low to high latitudes through the atmospheric wind systems as well as the ocean currents. Thus between the low to high latitudes a big thermodynamic engine is working with a source in the low latitudes and a sink in the high latitudes. This is what is known as the general circulation of ocean and atmosphere.

GAINS:

1. Direct radiation from the sun (Insolation), Q_S = 320 ly /day
2. Heat flow through ocean floor (0.1 ly/day)
3. Conversion of kinetic energy of waves into heat in the surf zone of coastal oceans
4. Heating through bio-chemical processes and nuclear reactions inside the oceans
5. Advective transfer of heat (rate of heat inflow due to currents), Q_V

All the gains from 2 to 4 are negligibly small and so are neglected.

LOSSES:

Heat exchange between the oceans and atmosphere through

1. back radiation, Q_B = 130 ly/day
2. Sensible heat transfer (conduction), Q_H = 20 ly/day
3. Latent heat transfer (evaporation), Q_E = 170 ly/day

Keeping all these points in mind, we can write the heat budget equation for any particular locality as

$$(Q_S + Q_V) - (Q_B + Q_H + Q_E) = Q_T$$

Where Q_T (balance) is the resultant rate of gain or loss of heat of a body of water at that locality. If $Q_T = 0$ means the temperature of the body of water is not changing. This does not mean that there is no heat exchange. It simply means that the net inflow equals the net out flow which is a steady state condition.

If we apply this to the world oceans as a whole

i) $Q_V = 0$ because all the advective flows are internal and must add up to zero as the average temperature of the world oceans remains constant.

ii) If we average over a whole year or a number of years the seasonal changes average out and Q_T becomes zero.

Then the heat budget equation can be written as:

$Q_S = Q_B + Q_H + Q_E$ Mosby has calculated the average values as

$320 = 130+20 + 170$ (which means 38% + 6% + 56% of Q_S). These values give some idea of the average magnitude of the heat budget terms.

4.1. DISTRIBUTION OF INSOLATION IN THE OCEANS (Q_S) :

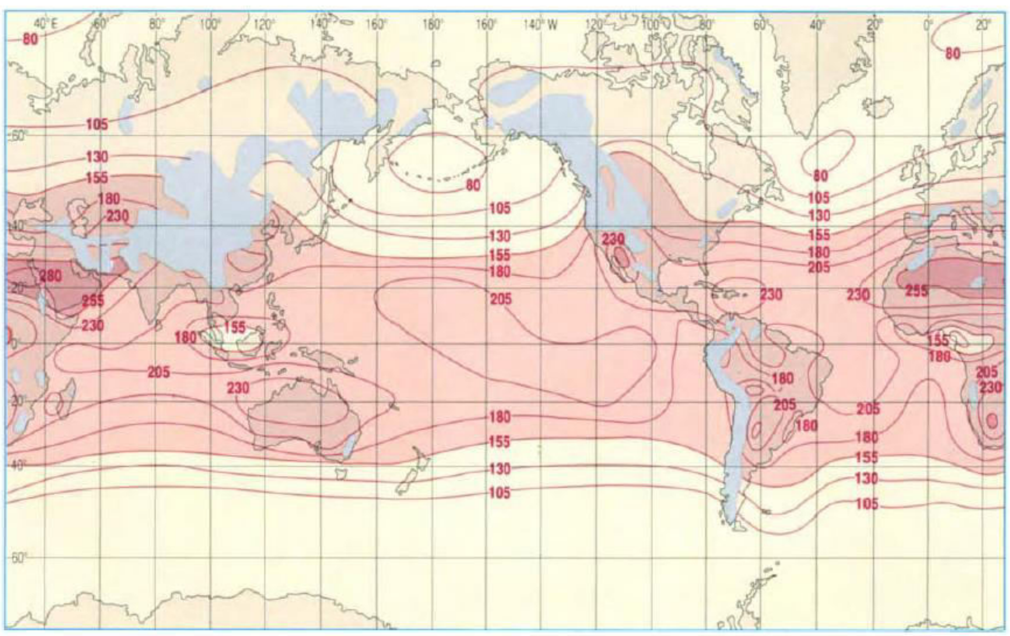


Fig.1. The amount of solar radiation received at the earth and ocean surface in J m^{-2} yr^{-1}

Fig.1 shows the variation of annual amount of solar radiation received over the ocean surface. The contours of Q_S appear to be parallel to the latitudes. In the tropical latitudes if Q_S between land and ocean over the same latitude is compared it is more over the land than the Oceans. The reason for this is the atmosphere over the oceans contains large amount of water

vapor which along with different gases like CO_2, SO_2 etc in the marine atmosphere absorb large amount of solar radiation (about 30%). Whereas the atmosphere over the land is drier and so penetration of insolation over land is more. However, the effect of water vapor in reducing insolation in low latitudes is greater than that in high latitudes because evaporation occurs at a greater rate from a warmer sea surface than a colder surface. Also as warm air can hold more moisture than the cold air, more water vapor is available in the low latitude atmosphere than the high latitude atmosphere. While the equatorial and tropical oceans receive about 205 J m^{-2} yr^{-1} it decreases to less than 80 J m^{-2} yr^{-1} in high latitudes in northern hemisphere. In some places (Malaysian and Indonesian region) the effect of water vapor is glaringly seen. While the insolation over the Indonesian region is 150 or so the adjacent East Indian Ocean area is over 200 J m-^{2} yr^{-1}.

4.1.1. Factors controlling Q_S:

The value of Q_S depends on a number of factors, i) the length of the day, season and the geographic latitude and longitude, ii) absorption of short wave radiation by gas molecules, dust particles, water vapor etc. present in the atmosphere iii) elevation of the sun's position and iv) cloud amount.

The effect of the cloud on insolation is given by the formula

$$Q_{SC} = Q_S (1 – 0.07 C)$$

where Q_{SC} is insolation reaching the earth's surface when 'C' amount of cloud is present,and Q_S is insolation with clear sky. For example if the sky is overcast, C = 10 then the insolation drop is 0.3 of Q_S which means 70% is lost.

A final factor that effects insolation is the albedo (34%) and the sea state which is determined by Beaufort wind scale.

4.2. BACK RADIATION (Q_B):

It is the effective long wave radiation given away from the ocean surface by virtue of emission of radiation as a black body that is governed by Stefan-Boltzman law. Back radiation depends on i) sea surface temperature ii) water vapor in the air and iii) cloud amount. The effect of cloud is given by the formula

$Q_{bc} = Q_b (1 – 0.08 c)$ where 'C' is the cloud amount. If the sky is overcast C = 10 and Q_{bc} will be 20% of Q_b. which means 80% is reduced.

Q_b also varies within the limits of 14% in tropical oceans. For example the variation of Q_b for the sea surface temperature (SST) of 10°C and 20°C can be computed with the help of Stefan-Boltzman law as

$Q_b = \sigma T^4$, therefore $Q_{b10} = \sigma (283)^4 = 225$ ly/day taking $\sigma = 8.22 \times 10^{-11}$

similarly $Q_{b20} = \sigma(293)^4 = 285$ ly/day. Therefore, $Q_{b20}/Q_{b10} = 0.14$ (=14%).

4.2.1. EFFECT OF WATER VAPOR ON Q_B:

With the increase of sea surface temperature the back radiation decreases. This is because a rise in SST causes an increase in the outward radiation from the sea but at the same time is accompanied by an increase in humidity as shown in fig. 2. The warmer the air over the oceans, the more water it can hold before becoming saturated. Thus a given relative humidity value at a high temperature corresponds to a greater atmospheric water vapor content than the same relative humidity at a lower temperature. The more water vapor is in the atmosphere, the longer wave radiation is absorbed by it, and the more is radiated back to the sea, thus decreasing the net loss of long wave energy from the sea surface.

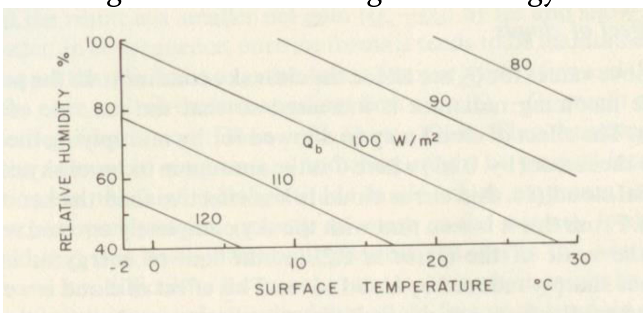


Fig.2. Back radiation (Q_b) from the sea surface as a function of SST and R.H in the absence of cloud

4.3. SENSIBLE HEAT TRANSFER (Q_H):

When there is a temperature gradient, the heat will be conducted from high temperature region to low temperature region. If the temperature decreases upward from the sea heat is conducted from the sea to the atmosphere and in such case Q_h is a loss term to the sea and if temperature increases upward, conduction takes place from the atmosphere to the sea and Q_h is a gain term to the sea. The amount of loss or gain is given by the equation:

Q_H is proportional to $C_p \dfrac{dt}{dz}$ or $Q_H = -KC_p \dfrac{dt}{dz}$

where 'K' is the coefficient of molecular thermal conductivity, C_p is the specific heat of air at constant pressure and dt/dz is the temperature gradient. The negative sign is assigned because the flow of heat will be towards the decreasing direction of temperature.

The coefficient of molecular thermal conductivity K, is a constant for a particular gas and for a particular temperature. However, in the applications of meteorological and oceanographical problems, particularly over the sea surface when the air is in motion it is turbulent. Hence the transfer of heat will be much more vigorous than the transfer of molecular motion. So in the case of turbulent motion the coefficient of molecular thermal conductivity is to be replaced by eddy conductivity of heat (A_h)

$$Q_H = - A_h C_p \frac{dt}{dz}$$

The value of A_h is not constant even at constant temperature. It depends on various factors such wind speed, size of the ripples waves, tides etc. in the oceans.

4.4. EVAPORATION (Q_E) :

The evaporation term Q_E is included in the heat budget because for evaporation to occur it is necessary either to supply heat from outside or heat is to be taken from the remaining liquid.

$Q_E = (F_E \cdot L_T)$ where F_E is the rate of evaporation of water in grams per minute per square centimeter and L_T is the latent heat of evaporation in cal/gram. L_T is not constant and depends on temperature. For pure water L_T is given by:

$L_T = 596 - 0.52\,t$ where 't' is the temperature in ^{O}C

F_E is proportional to $\frac{df}{dz}$, $F_E = -A_E \frac{df}{dz}$

where $\frac{df}{dz}$ is the humidity gradient and A_E is eddy coefficient of diffusion which depends on wind speed, size of the ripples and waves.

$$\therefore Q_E = F_E . L_T = -L_T . A_E \frac{df}{dz}$$

4.5. BOWEN'S RATIO:

Because of the uncertainty in the values of A_h and A_E, it is not possible to compute Q_H and Q_E directly. Bowen introduced a method by which Q_H and Q_E can be evaluated from the heat budget equation: $Q_S = Q_b + Q_h + Q_E$

In this Q_S and Q_B can be measured using the pyranometer and radiometer.

$$Q_S - Q_B = Q_H + Q_E$$

$$(Q_S - Q_B)/Q_E = \left[\frac{Q_H + Q_E}{Q_E}\right] = 1 + \frac{Q_H}{Q_E} = 1 + R$$

$$\therefore Q_E = \left(\frac{Q_S - Q_B}{1+R}\right)$$

$$or Q_H = \left(\frac{Q_S - Q_B}{1 + \frac{1}{R}}\right) \quad \text{................ (4.1)}$$

This R is called Bowen's ratio. If we can determine R, either Q_H or Q_E can be determined in an indirect way. As quantifying Q_H and Q_E is difficult, they can be estimated if we can estimate R and R can be estimated using the vertical gradients of heat ($\frac{dt}{dz}$) and

water vapor $\left(\frac{df}{dz}\right)$ as given below. Also as long as radiation terms (Q_H, Q_E) or net radiation term (Q_S-Q_B) is available we can estimate R.

Substituting the values of Q_H and Q_E in R $= \frac{Q_H}{Q_E} = \frac{-A_H C_P \frac{dt}{dz}}{-L_T A_E \frac{df}{dz}}$

A_H and A_E being the coefficient of eddy thermal conductivity and coefficient of eddy diffusivity. Hence under similar conditions the numerical value of these two terms can be assumed to be same values. So $A_H = A_E$

R $= \frac{C_p \frac{dT}{dz}}{L_t \frac{df}{dz}}$ if $\frac{df}{dz}$ is written in terms of vapor pressure units using the formula

$\frac{df}{dz} = \frac{0.621}{p}\frac{de}{dz}$ where p is the atmospheric pressure expressed in millibar and $\frac{de}{dz}$ is the vapor pressure gradient.

$\therefore R = \frac{p}{0.621}\frac{C_P}{L_t}\frac{dt/dz}{de/dz}$ if we assign suitable temperatures and vapor pressures within a sufficient layer of thickness between the heights Z_1 and Z_2 $\frac{dT}{dz} = \frac{T_S - T_a}{Z_1 - Z_2}, \frac{de}{dz} = \frac{e_s - e_a}{z_1 - z_2}$

$\therefore R = \frac{p}{0.621}\frac{C_P}{L_t}\frac{T_S - T_a}{e_S - e_a}$ Substituting standard values of p=1000 mb, C_P =0.24 and L_t = 580 cal, $R = 0.66\frac{T_S - T_a}{e_S - e_a}$

By measuring the temperatures and aqueous vapor tension at the sea surface and anemometer level (ten meters) above the sea surface R can be calculated. Thus Bowen's Ratio is defined as the ratio of sensible heat to latent heat loss.

The average value of R in the tropical oceans is about 0.1 which implies that the latent heat loss in the oceans is ten times greater than the sensible heat loss. This value increases to about 0.45 at 70^{O}N latitude. Which means R increases with increase of latitude implies Q_H proportion increases from Q_E in the Bowen's Ratio. Lower the values of R, the higher will be the evaporation. Usually R will be positive because the average sea surface temperature is higher than the average air temperature by about 0.8^{O}C.

A negative value of R indicates gain of heat by the oceans which can occur when the sea surface temperature is lower than the air temperature or when condensation takes placing on the sea surface.

The value of R will be the lowest in the subtropical regions in winter particularly on the western sides of the oceans where maximum evaporation takes place. Towards the

equator the value of R increases slightly. The higher values of R occur during summer in high latitudes which imply that sensible heat loss is much more there compared to low latitudes.

4.6. Latitudinal variation of Heat Budget:

The Fig 3b represents the net gain of heat in the equator ward from 23^{O}N and a net loss poleward. Another important point to be borne in mind with respect to Fig.3b is the hatching area above and below the mean indicates the heat gain and loss on either side of 23^{O}N respectively is equal which means heat gain is equal to heat loss.

This balance can be explained only by means of general circulation of the oceans. That is the advective term (Q_V) in the heat budget term is to be considered important. Warm currents flow towards high latitudes and cold currents flow towards low latitudes for transport of heat or cold. Thus over a period of year the mean temperature of the oceans remains constant. Note that 23^{O}N is the demarcation latitude between gain and loss in the oceans as per Fig.3b. At this latitude (23^{O}N) mean Q_T =0 and on either side of it the lower latitudes experience gain and higher latitudes experience loss in each hemisphere.

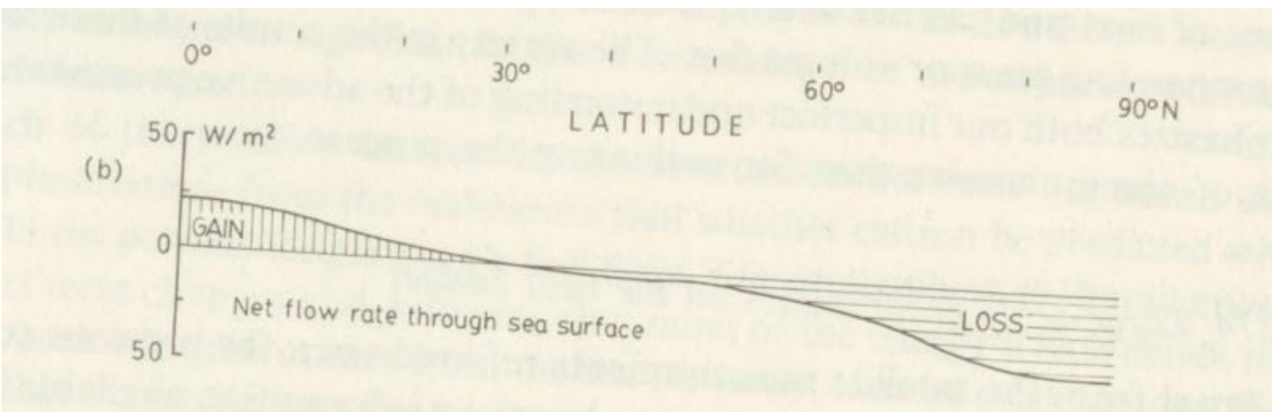


Fig.3.The: Net balance of heat gain to loss in N.H. Note the balance on either side of latitude 23^{0}N

4.7.1. Quantitative net Global heat balance of oceans:

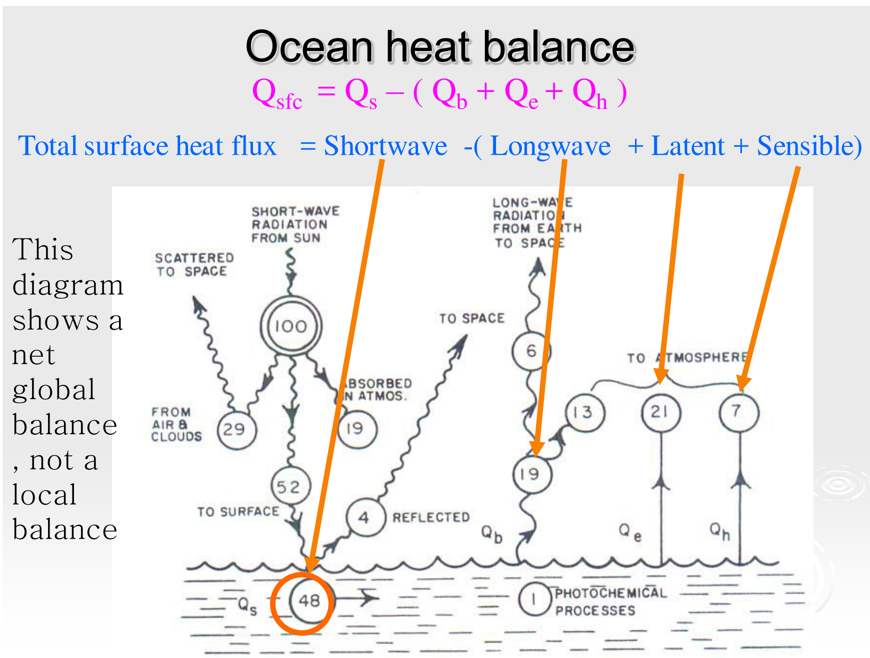


Fig.4. Net Global heat balance of oceans

From figure 4:

Short wave balance:
Sun is supplying 100 units. Out of the 100 units supplied by sun, 52 are directly going to oceans. Out of 52, 48 units are absorbed by oceans and 4 units are reflected to space.
Atmosphere is also directly receiving 29 and19 units from sun. The 29 units are scattered to space and the 19 units are absorbed by the atmosphere. So 29+19+52 (48+4)=100 .

Long wave balance:
Out of the 48 units absorbed by the oceans, 1 unit is used for photochemical processes, 19 units lost as back radiation, 21 units lost as evaporation and 7 units lost as sensible heat transfer. Out of the 19 units of back radiation of earth 13 units are going to atmosphere and 6 units are going to space.
1+19 (13+6) + 21 + 7 = 48

Balance of oceans:
Sun's supply to oceans : 52; Loss of oceans : 48; Total : 52 + 48 = 100

HANDOUT OF ARBAMINCH UNIVERSITY

INSTRUCTOR: Prof.(Dr.)A.S.N.MURTY

OCEANOGRAPHY & MARINE METEOROLOGY

MHD 0904

B.Sc. II Year: II Semester

Chapter V

Ocean currents & Water masses

5.1. Ocean currents:

Oceans are not just bodies of stagnant water. Ocean water moves from one place to another. Waters at different depths move in different directions and at different speeds. Currents are generated due to wind and density difference. Most of the surface currents are driven by wind and subsurface currents are driven by density difference. The subsurface circulation is called thermohaline circulation. A modified version of surface and subsurface circulation together is called conveyor circulation as shown in Fig.1.

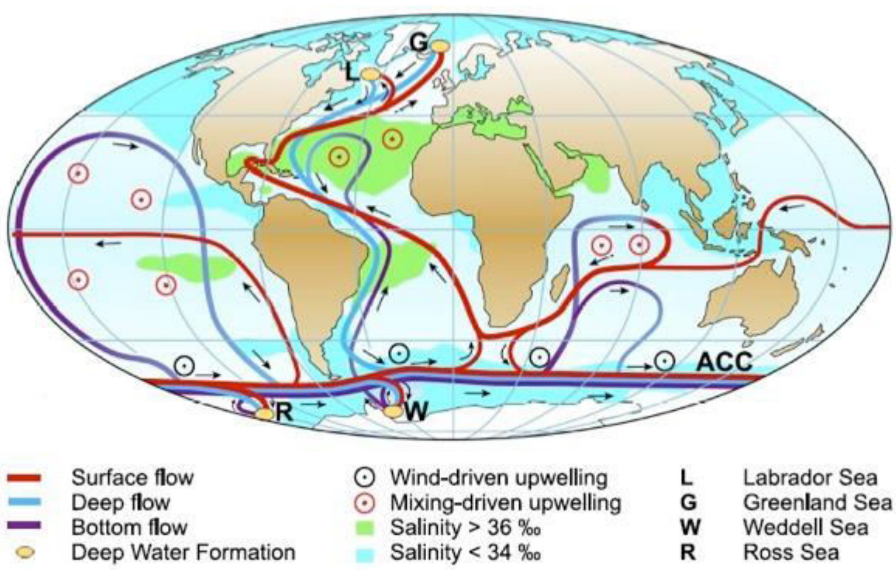


Fig.1. Thermohaline & conveyor circulation of global oceans. Light curve shows bottom flow and dark curve shows surface flows.

The cold water in Green land (G) and Labrador (L) and Weddell (W) and Ross(R) seas sink at G L,W and R along the route shown by blue light curve and returns back from surface along the thick red curve. That is why this is called conveyor circulation.

Different names of currents of the world oceans along with warm and cold currents are shown in Fig.2 & 3. Generally warm currents flow towards the poles and cold currents flow towards the equator. The names of different major currents are indicated in the Fig.2 as per the table 1.

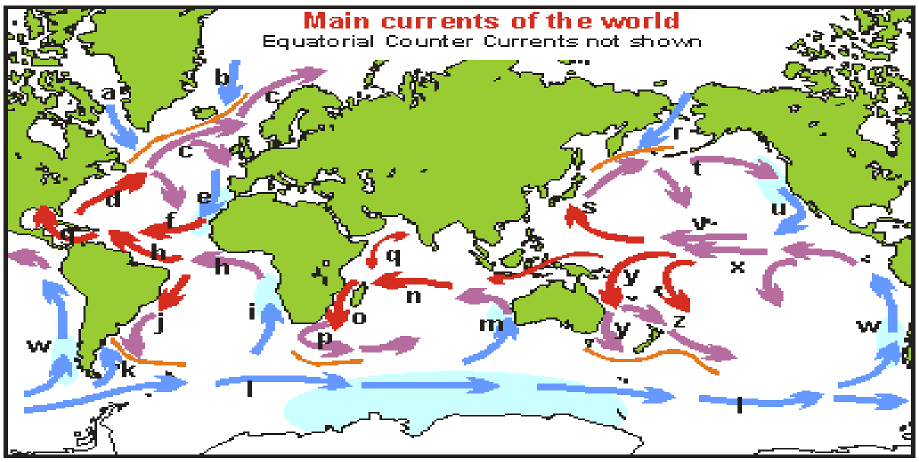


Fig.2.Major currents in the oceans

Table 1 Names of Major current systems of the world oceans as in the Figure 5.1.

a Labrador Current	n South Equatorial C.
b East Greenland C.	o Mozambique C.
c North Atlantic Drift	p Agulhas C.
d Gulf Stream	q Monsoon Drift
e Canary C.	r Kamchatka C./Oya Shio
f North Equatorial C.	s Kuro Shio C.
g Caribbean C.	t North Pacific Drift
h South Equatorial C.	u California C.
i Benguela C.	v North Equatorial C.
j Brazil C.	w Peru/ Humboldt C.
k Falkland C.	x South Equatorial C.
l West Wind Drift	y East Australia C.
m West Australian C.	z East Auckland C.

5.1.1. Major features of ocean circulation:

The major features of the general circulation of the oceans comprise of gyres, Equatorial current system, and western and eastern boundary currents. However, a clockwise circulation in the Northern Hemisphere (N.H) and an anti- clockwise circulation in the Southern Hemisphere (S.H) are found in all major oceans called gyres as shown in Fig 3. These gyres are called the back bone of the circulation. The

equatorial current system comprises of westward flowing North and South equatorial currents, eastward flowing equatorial counter current, equatorial jets and eastward flowing equatorial under current. While western boundary currents are narrower and swifter, eastern boundary currents are generally weak and diffused. This strong western boundary currents are called "westward intensification" Indian Ocean circulation is partly different from that of the other major oceans.

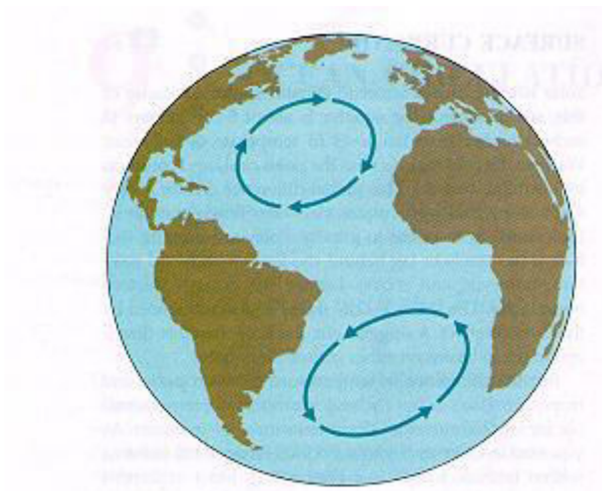


Fig.3. The clockwise and anticlockwise gyres in the northern and southern hemispheres- the back bone of the circulation.

The general circulation showing gyres, warm and cold currents in all the oceans are shown in Fig.4. The arrows tending toward poleward are warm and arrows tending toward equatorward are cold currents.

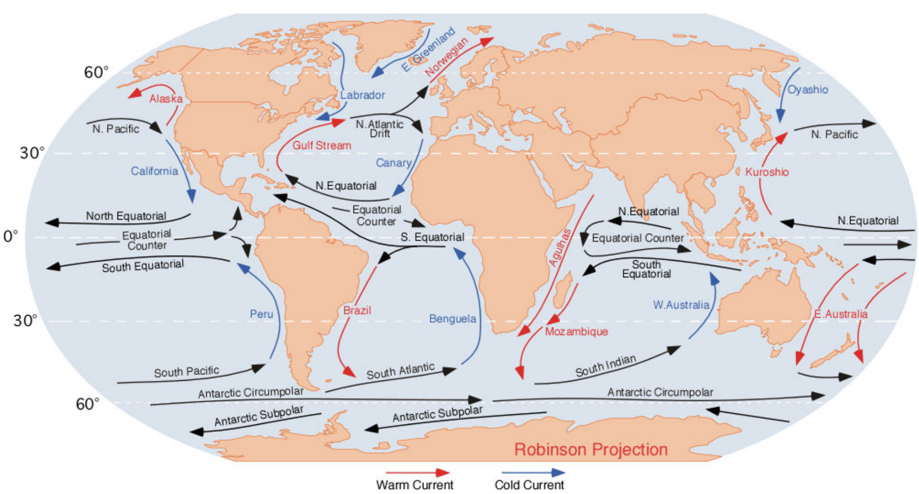


Fig.4. General Circulation of the oceans along with gyres.

5.2. TEMPERATURE- SALINITY DIAGRAM (TS-DIAGRAM):

Helland-Hansen pointed out in 1916 when in a given area the temperatures and corresponding salinities of the sub surface water are plotted against each other the points generally fall on a well defined curve called TS-curve, showing the temperature salinity association of that water. The TS-diagram containing salinity on the x-axis, temperature on the y-axis and σ_t lines over printed on it is called TS-diagram as shown in Fig.5. Montgomery used specific volume anomaly lines or thermosteric anomaly lines in place of σ_t lines and called it as TSV-diagram.

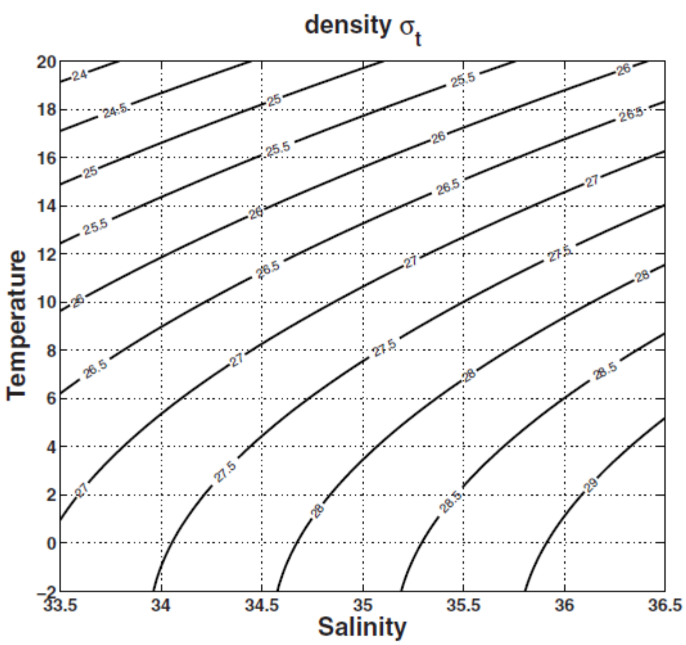


Fig.5. TS-diagram

5.2.1. USES OF TS-DIAGRAM:

The TS-diagram has become one of the valuable tools in physical oceanography. By means of this diagram characteristic features of the temperature salinity distribution are conveniently represented and anomalies in the distribution are easily recognized. The diagram is also useful in correcting the data. More important use is water masses can be classified on the basis of temperature-salinity characteristics. But density cannot be used for classification because two water masses of different temperatures and salinities may have the same density. In recent years the amount of mixing between one water mass to the other is also being estimated with the help of TS-diagram.

A water mass is defined by a portion of Ts-curve on the TS-diagram which has a range of temperature and a range of salinity of a huge body of water of almost homogeneous characteristics whereas a water type is represented by a point on the TS-diagram which has a single value of temperature and a single value of salinity and has no thickness.

5.3. FORMATION AND MOVEMENT OF WATER MASSES:

Generally water masses are formed due to stagnation in a particular locality for a long time. Usually this happens in the areas of convergence zones as water stays for a long time in those regions. When they stay there for a long time the climatological character of that region is gained by that water due to conduction and convection. After the formation, the water mass spreads at a level determined by its density along the path of its σ_t surface and occupies the subsurface depths in the lower latitudes as shown in the Fig.6. Some waters also form locally and spread only to short distances. Usually most of the water masses are formed at subtropical and Antarctic convergence zones.

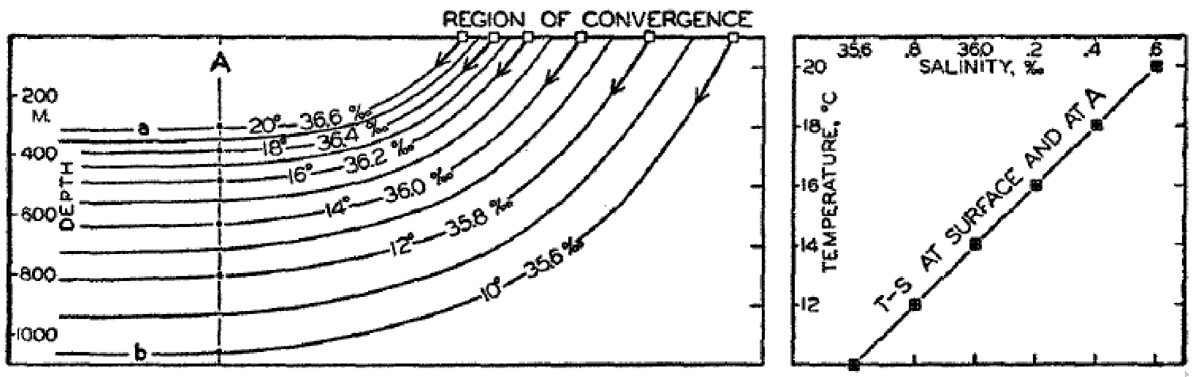


Fig.6. Schematic representation of the formation of a water mass at the convergence and by sinking along Sigma-t surfaces. The diagram to the right demonstrates that the vertical T-S relation of the water mass agrees with the horizontal T-S relation at the surface in the region of convergence.

Though water masses flow by sinking along σ_t (Sigma-t) surfaces, they do not readily mix and will show a range of temperature and salinity (10 to 20^0C and 35.6 to 36.6 ‰) as indicated in T-S relation between the lines 'a' and 'b' in Fig.6 (the vertical line A).

The Fig.7 indicates the approximate locations and boundaries of the different surface water masses in the three major oceans of the world. The central water masses are formed and spread at the subtropical convergence, while the Antarctic Intermediate water is formed and sink at Sub Antarctic convergence and the Antarctic bottom water is formed and spread from the sills of Antarctic continent. The water masses are named after the name of the locality of the ocean they either form or exist.

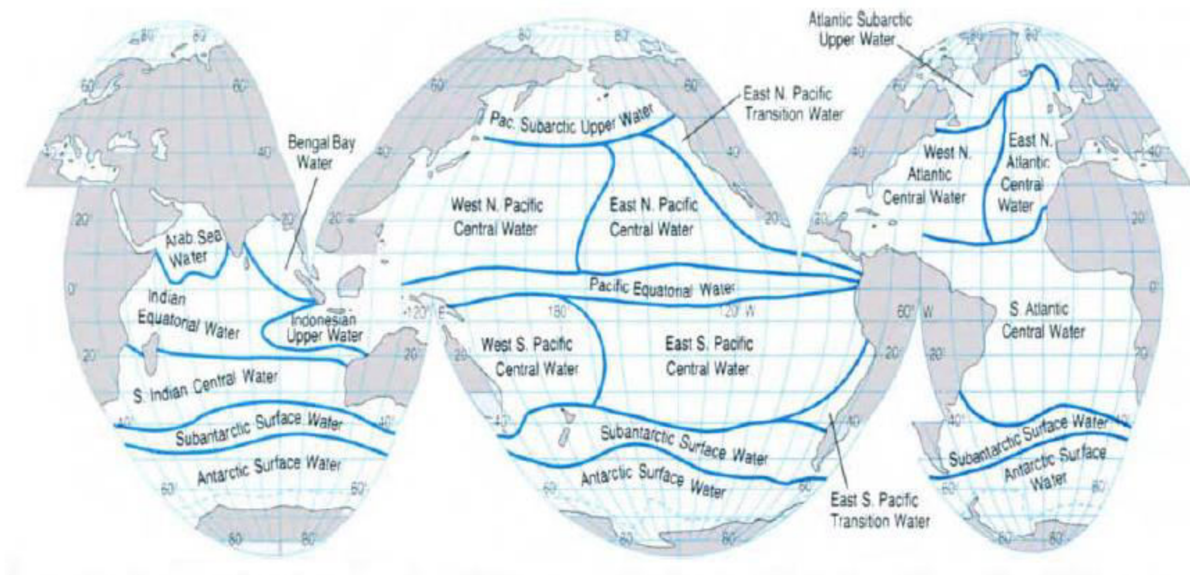


Fig.7. Approximate locations, boundaries and names of the surface water masses of the Oceans.

Figure 7 shows the following water masses at the sea surface in all the three major oceans.

Zone I (Indian Ocean) :

1. Antarctic Surface Water, 2. Sub-Antarctic Surface Water, 3. South Indian Central Water, 4. Indian Equatorial Water, 5.Arabian Sea Water, 6. Indonesian Upper water, and 7. Bay of Bengal Water.

Zone II (Pacific Ocean):

1. Pacific sub arctic upper water, 2. West North Pacific Central Water, 3. Eastern North Pacific Central Water, 4. Pacific Equatorial water, 5. Eastern North Pacific Transition Water, 6. West South Pacific Central water, 7. East South Pacific Central water, 8. Sub Antarctic Surface water, 9. Antarctic Surface Water, 10. East South Pacific Transition water.

Zone III (Atlantic Ocean):

1. Atlantic Sub Arctic Upper Water, 2. West North Atlantic Central Water, 3. East North Atlantic central water, 4. South Atlantic Central Water 5. Sub Antarctic suface water 6.Antarctic surface water.

HANDOUT OF ARBAMINCH UNIVERSITY

INSTRUCTOR: Prof.(Dr.)A.S.N.MURTY

OCEANOGRAPHY & MARINE METEOROLOGY

MHD 0904

B.Sc. II Year: II Semester

Chapter-VI: Coastal bodies

7. ESTUARIES

An estuary is a **semi-enclosed coastal body of water** that has **free connection to the open sea** and **within which sea water is measurably diluted with fresh water derived from land drainage**. A typical estuary has most of the fresh water being discharged at its head (river joining place) and has a transitional section (mouth) between the body of the estuary and the coastal ocean as illustrated in Fig.1.

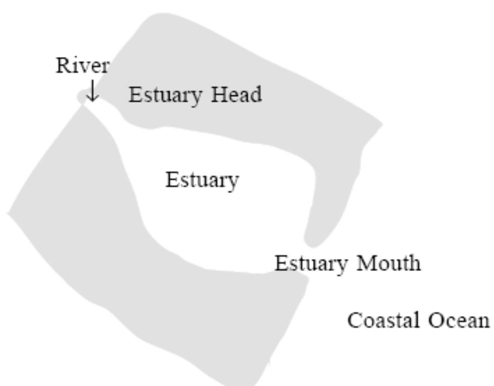


Fig.1. A typical sketch of an estuary

The typical circulation in estuaries is shown in Fig.2. P_4 is the line of Level of no motion or it is the demarcation line between the less saline riverine water above and more saline sea water below.

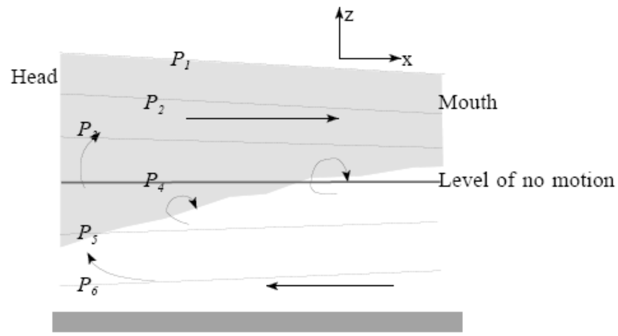


Fig.2. Circulation in an estuary

7.1. Classification of Estuaries:

Estuaries are classified basically in four ways: Geology, water balance, salinity structure and circulation.

1) Geomorphology is based on the geologic origin of the basin. 2) Water Balance is based on the balance between water input like (precipitation, river discharge, thawing) and outputs like (evaporation and freezing). 3) Salinity Structure - based on the water column stratification 4) Stratification- currents and Circulation.

7.1.1. Classification of estuaries based on their Geomorphology

Based on their geomorphology, most estuaries can be identified within one of the following four categories:

i) Coastal plain or drowned river valley - inundated 7,000 years ago, when sea level rose after the Pleistocene Ice Age (15,000 years ago). Examples: Chesapeake Bay, Delaware Bay, Hudson River Estuary in U.S.A.

ii) Fjords – Over deepening by active glaciations. They have old glaciers that form sills near the mouth of the estuaries. Examples: Juan de Fuca Strait, Puget Sound, Fjords in Norway, British Columbia and Chile.

iii) Bar-built - these are lagoons developed on low coastlines that have narrow passages connecting to the sea (inlets). Examples:Krishna, Godavari and Mahanadi estuaries & Chilaka Lake.

iv) Tectonic - Formed by tectonic faults, land slides, or volcanic eruptions. Example: San

Francisco Bay.

7.1.2. Classification of estuaries based on their Water Balance

i) Positive estuaries - freshwater input exceeds evaporation,

ii) Negative or inverse estuaries – Fresh water loss exceeds the input, e.g., Mediterranean Sea, Red Sea.

7.1.3. Classification of estuaries based on their Salinity Structure

This classification takes into account mean salinity structure of the estuary basing on the flow characteristics in and out of the estuary. It is not exact because many estuaries will change from one type to another in a matter of days or months so use this classification cautiously.

i) Vertically mixed or vertically homogeneous - isohalines are straight from surface to bottom along the estuary. If the estuary is wide enough, net outflow will develop from surface to bottom in parts of the cross-section, and net inflow will appear from surface to bottom in other regions. If the estuary is too narrow for the transverse partition of inflows/outflows to develop, then net flow will be seaward from surface to bottom. Mass balance will be achieved through landward mixing of salt. Mixing (destratifying) processes tend to dominate over stratifying tendencies from river discharges.

ii) Weakly Stratified or partially mixed - Vertical stratification is apparent throughout the water column. Inflow and outflow volumes are similar. Mixing completes equally with buoyancy tendencies or tidal prism (volume of estuary between high and low water levels) is comparable to freshwater volume.

iii) Strongly or highly stratified - Stratification is better developed than in the partially mixed estuaries. Buoyancy tendencies dominate over mixing tendencies.

iv) Salt wedge estuary – This type of estuary develops where a river discharges into a virtually tide less sea as shown in Fig.3a. In this estuary less dense river water spreads above the surface of denser saline sea water. Between the fresh water and sea water, there are very sharp density and salinity gradients so that a stable halocline develops and the two water types do not mix easily (Fig.3a & b). However, because one layer of water is moving over another layer, shear stress occurs at the interface producing turbulence at the interface of the two layers. This results in generating series of internal waves at the interface.

The position of the salt wedge is dependent on the river flow. When the discharge of fresh water is low, the salt wedge can penetrate further in land than when the discharge is high. Only rivers with very low rates of sediment discharge form open salt wedge estuaries. If the discharge of sediment is high, then it tends to accumulate to build a delta. This is the case in the case of Ganges, Godavari and Krishna estuaries in India.

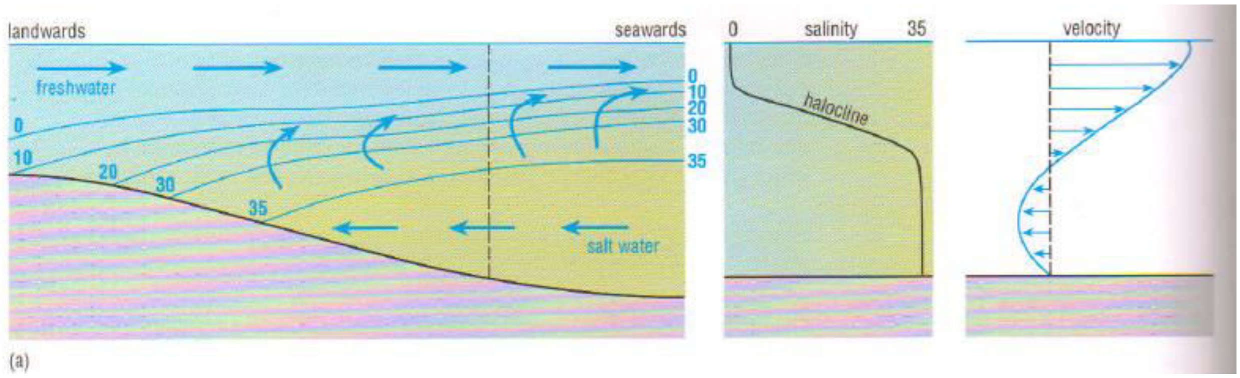


Fig.3a. circulation, salinity distribution and velocity gradients in a salt wedge estuary. The flow of fresh water at surface is towards the sea and the salt water towards land at the bottom. Vertical mixing takes place at the interface. The halocline profile in the second figure is at the position indicated by dashed line in first figure. Third figure is velocity-depth profile along the dashed line showing the residual flows.

Partially mixed estuary:

These estuaries occur where rivers discharge into the sea with a moderate tidal range. Tidal currents are significant and so the whole water moves up and down the estuary with the flood and ebb tides. Consequently, in addition to the current shear at the interface, friction at the estuary bed creates shear stress and generates turbulence which causes even more effective mixing of the water column than that caused by waves at the interface of fresh water/ salt water. Not only the salt water mixed upwards, but fresh water is mixed downwards as shown in Fig.3b. This two way mixing makes the halocline less well developed.

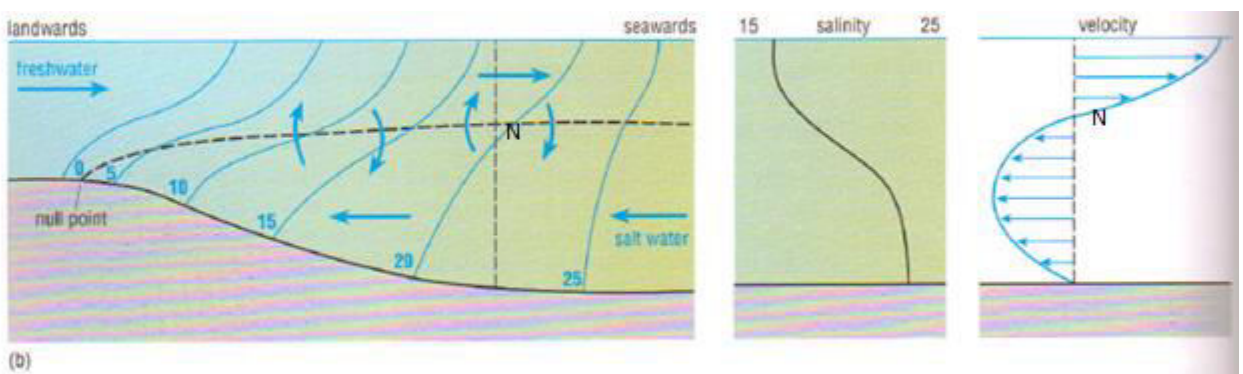


Fig.3b. circulation and salinity and velocity in a partially mixed estuary. The dashed horizontal line in first figure indicates the depth at which no horizontal residual flow occurs either land or seaward. Second figure is poorly developed halocline at the vertical dashed line. The velocity profile in third figure indicates a marked upstream (landward) residual flow of sea water at the bed. The point 'N' is null point where no net land or sea ward flow exists.

The fresh water flowing sea wards is now mixed with a relatively high proportion of salt water so that the compensating land ward flow from the sea is much stronger than in the salt wedge estuary. These upper seaward and lower landward flow from the sea distort the isohalines in a longitudinal section of the estuary as shown in first figure of 3b.

The currents caused by the mixing of fresh water and salt water in salt wedge and partially mixed estuaries are referred to as residual currents as they are less than 10% of tidal currents. They are important in understanding the transport of sediment.

Towards the head of the estuary net landward bottom flow diminishes and net seaward upper flow increases. The depth at which no net landward or seaward movement of water exists is called null point (N in third figure of 3b).

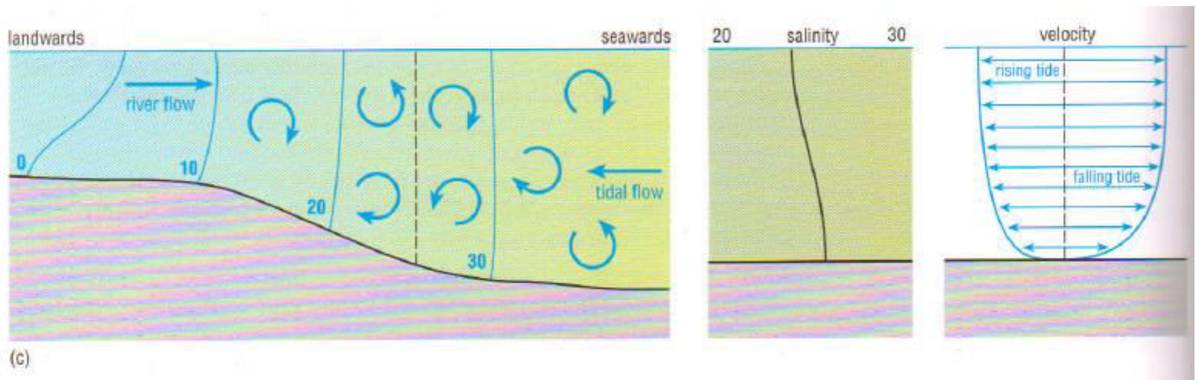
Residual or net flows (horizontal arrows) in 3 (a) and (b) are seawards at the surface because of the river flow, landwards at the bottom because of vertical mixing and entrainment across the river water /sea water interface.

In 3(b), the dashed sub-horizontal line on the longitudinal section shows the depth at which no horizontal residual flow occurs either seawards or land wards and its intersection with the bed near the head of the salt water intrusion defines the null point. Although this null point is shown only in 3(b), it occurs in any tidal estuary.

Partially mixed estuaries are common in East coast of India. They are Ganges, Brahmaputra, Godavari, Krishna etc.

Well mixed estuary:

Because of strong tidal currents and Swift River flow into the estuary the water is well mixed. In the well mixed estuaries, salinity hardly varies with depth while it varies considerably in the horizontal. The coriolis force tends to swing the incoming tidal flow and seaward flowing river water to the right hand side in Northern hemisphere and to the left in the Southern hemisphere. This means that in N.H the sea water flows up to wards the land on the left side and river water flows towards sea on the right hand side. While moving like this some mixing laterally takes place so that a horizontal residual circulation is developed. A poor halocline is developed and the currents strong and decrease uniformly with depth as shown in third figure 3c. Net flow in 3(c) is landwards on the flood tide, seawards on the ebb.



The Fig.3c. represents the water circulation , salinity distribution and velocity gradient within the well mixed estuaryThe dashed vertical line is the area where the salinity and velocity profiles are shown. Note the progressive weakening of the halocline from Figures 3(a) to (c). This weakening is due to progressive increase of tidal forcing. In well mixed estuary (c), the salinity of the

water column at any particular point in the estuary depends up on the state of the tide. Curved arrows on the longitudinal sections represent mixing.

Seiches

A Seiche is a standing wave, which can be considered as the sum of two progressive waves travelling in opposite directions as shown in Fig. 4. Seiches can occur in Lakes, bays, estuaries and harbours which are open to the sea at one end.

In most of the coastal bodies like bays and estuaries, the water is relatively shallow compared to the wavelength of the seiche (L) and the period of the Seiche is determined by the length of the basin and the depth of the water. Figure 4a &b shows the idealized water motions in a seiche of a closed basin.

At either end of the container, water level is alternately high and low, whereas in the middle the water level remains constant. The length of the container (l), corresponds to half the wavelength (L) of the Seiche in Fig.4a.

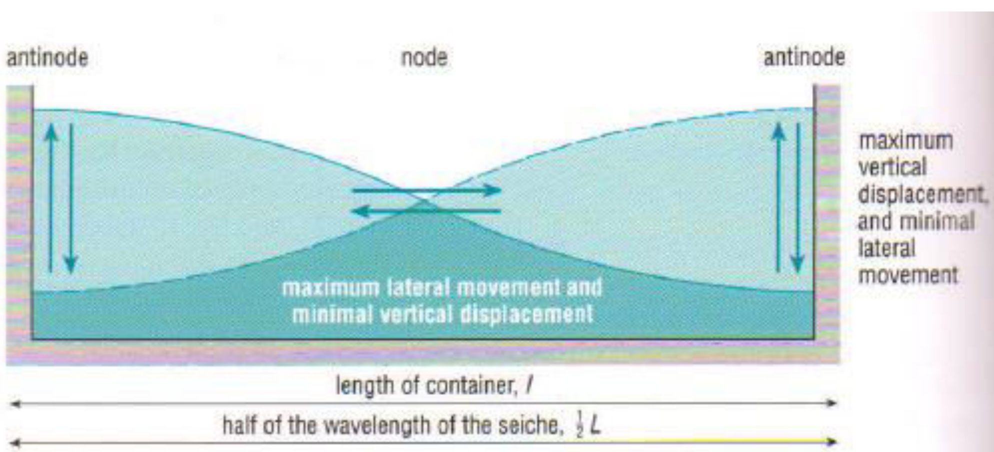


Fig.4a. Idealized profile of a seiche in a closed basin

The above figure 4a is an idealized vertical profile of a seiche. At either end of the basin alternately the water level rises and falls, while at the middle the water level remains constant. The length of the basin (l) corresponds to half the wave length (L) of the seiche. While at node the horizontal flow is maximum, at antinode the vertical flow is maximum.

If the water depth divided by the length of the basin is less than 0.1, then seiche can be considered as a shallow water wave. Then its period of oscillation (vertical movement at anti node) is:

$$T = \frac{2l}{\sqrt{gd}}$$

where l = length of the basin, d = depth of the basin, g = 9.8m/sec^2

In most bays and estuaries, the water depth is relatively shallow compared to the seiche wave length (L) and the period of the seiche is determined by the length of the basin (l) and water depth (d).

In some basins open to sea at one end as shown in Fig.4b, the node can occur at the entrance to the basin and an antinode at the landward end. In this case, the length of the basin (l), corresponds to a quarter of the wave length of the seiche (L). Then the period of oscillation of seiche is:

$$T = \frac{4l}{\sqrt{gd}}$$

Here 'T' is also known as the resonant period. For standing waves to develop, the resonant period of the basin must be equal to the period of the wave or to a small whole number of multiples of that period as:

$$T = \frac{4l}{n\sqrt{gd}}$$

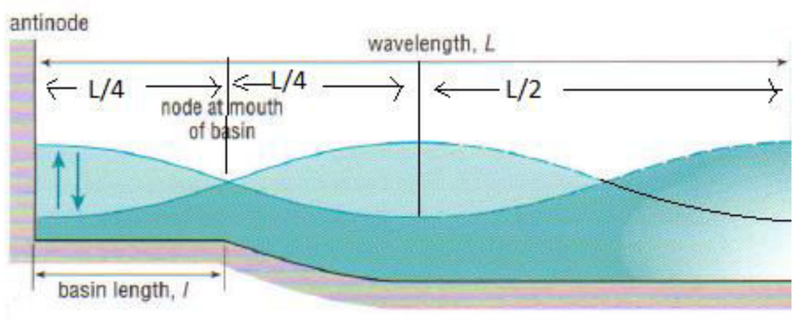
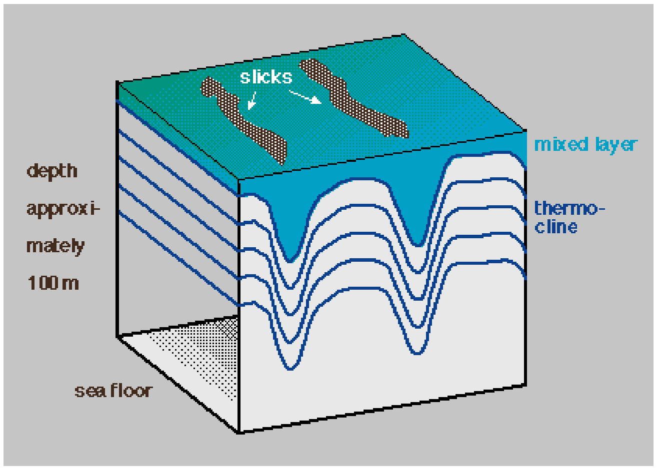


Fig.4b.Idealized wave profile of a seiche in a open end

Internal waves

Sometimes these gravity waves can be generated at an interface between two layers of ocean water of differing densities. These waves occurring at subsurface where there is change of density are called internal waves. These generally occur at pycnocline, where there is a rapid increase of density with depth. In other words they occur at thermocline or halocline as density depends on temperature and salinity. Internal waves travel considerably more slowly than most surface waves. They have greater amplitudes, longer periods and longer wavelengths than that of the surface waves. Internal waves are of considerable importance in the context of vertical mixing processes in the oceans, especially when they break. The occurrence of internal waves can be detected from surface by the presence of slicks that are

formed due to the convergence of water above the internal wave troughs in the mixed layer as seen in Fig.5.



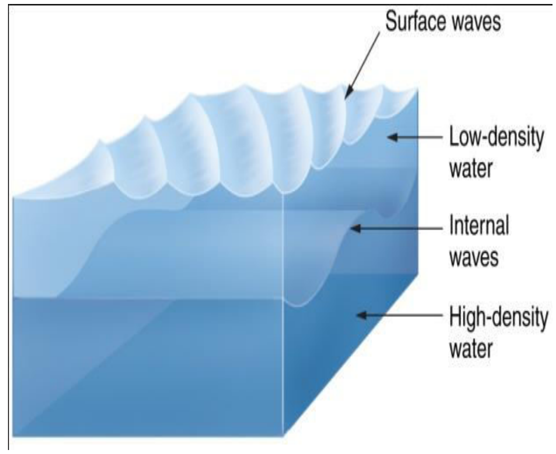


Fig.: Sketch of an internal wave propagating at the seasonal thermocline in the coastal ocean. The curved lines indicate isotherms (contours of constant temperature). The mixed layer is indicated by the greenish-blue region above the first isotherm. The crowded and curved troughs of isotherms at mid-depth indicate the thermocline. The slicks at the surface are produced by the convergence of water above the wave troughs in the mixed layer. The depth range of occurrence of internal waves may be around 100 m.

Internal waves are also called 'dead water' as they reduce the speed of moving ship when the ship moves against the direction of the ship as shown in Fig.6.

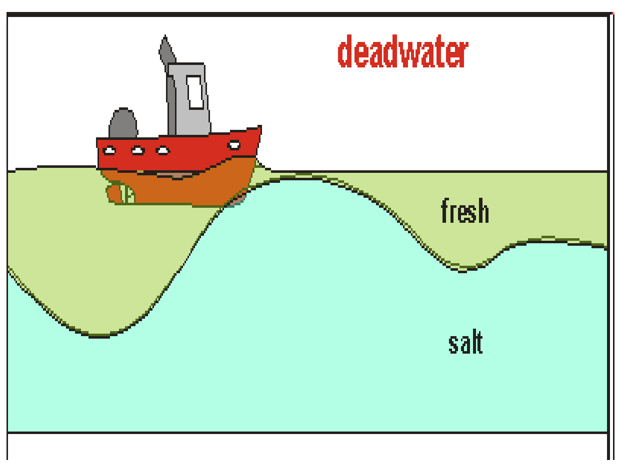


Fig.6 shows the prevention and reducing the speed of moving ship by internal wave called dead water.

INSTRUCTOR: Prof.(Dr.)A.S.N.MURTY

OCEANOGRAPHY & MARINE METEOROLOGY

MHD 0904

B.Sc. II Year: II Semester

Chapter-7: Waves & Tides

7.1. Wave Basics:

The wind blows over the water, changing its surface into ripples and then to waves. Thus most of the normal waves are wind-driven. As waves grow in height, the wind pushes them along faster and higher. Sometimes waves can become unexpectedly strong and destructive. As waves enter shallow water, they become taller and slow down, eventually break on the shore. Waves cut the rocks and make them erode faster and create beaches by transporting sand towards the shore. Waves are only oscillations in the water's surface and don't transport much of mass. Waves are Movement of energy along air-sea interface.

Waves are defined as successive crests with intervening troughs advance in an undulatory motion.

Waves are considered under two categories. They are forced waves and free waves. A forced wave propagates as long as the generating force acts and once the generating force is removed the wave stops whereas a free wave continues to propagate even after the removal of the generating force.

Waves can be distinguished basing on the shape of the profile. If the profile of a wave moves relative to the medium, it is a progressive wave. If the profile doesn't move, but merely oscillates in one place relative to the medium, it is a standing wave. A wave in which the particle motion is entirely parallel to the direction in which the wave moves is called a longitudinal wave. One in which the motion is entirely perpendicular is called a transverse wave. Ocean surface waves belong to neither of these classes as these are theoretical profiles and as ocean waves are formed due to combination of several profiles and bands called spectrum.

While the crest of the wave is the highest point of water above the mean water level, the trough is the lowest point below the water level. In the case of sinusoidal wave the crest and the trough are displaced symmetrically from the mean level and so the height of the wave is twice the amplitude or crest height as shown in Fig.1. The wavelength,L, is the horizontal distance between two crests and the period,T, is the time interval between the passage of two successive crests past a fixed point. The wave velocity or Celerity,C, is the ratio of wave Length to time period (L/T).

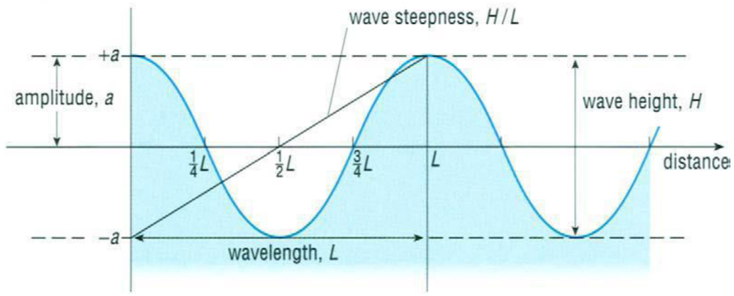


Fig.1.Wave profile showing the different wave parameters

7.2. Wave generation by wind:

In the Ocean waves are generated due to various forces. The normal day to day waves are generated due to wind and so they are called wind waves. The restoring force for *small amplitude* waves is either surface tension or gravity and so they are also called *gravity waves*. But for the *finite or nonlinear* waves the restoring force is gravity.

The region of blowing wind is called *generating area.* The generating area is highly chaotic and confusing and contains high energy as shown in Fig.2.The development of waves ultimately depends on three factors *speed, duration* and *fetch* of the wind. The duration of time that the wind acts on the water surface is called the wind duration. The distance over which wind blows called the fetch. A minimum speed, duration and fetch are necessary for the development of wavelets. When these minimum conditions over steady state reach then the sea is called *fully arisen sea*.

The first wavelets appear in the fully arisen sea are *capillary waves* which are only a few centimeters in length and a few meters in height. The capillary waves, though small, wrinkle the sea surface. This wrinkling is restricted by the surface tension. However, if the wind continues to blow, the capillary waves grow in size and become gravity waves.

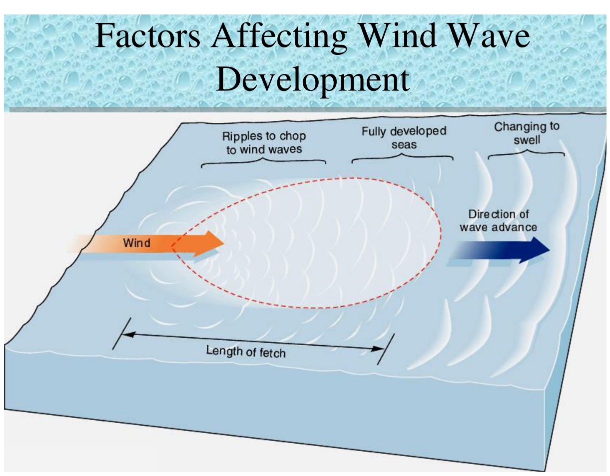


Fig.2: Generating area and wave generation mechanism

7.3. **Seas and Swells**:

Waves that are generated in the generating area are called *seas.* All the seas that are generated in the generating area gets nullified sometimes and some resulting waves will come out of this generating area and can travel to far off distances up to the coasts. These long period waves that come out of the generating areas and reach the coasts are called *swells*.

Sea waves are usually shorter period (higher frequency) than *swell*. Generally 10 seconds (0.1 cycles per second) can be taken as the demarcation period from sea to swell. *Sea* is shorter in length, steeper, and more rugged and confused in state than the swell as shown in the Fig.3.

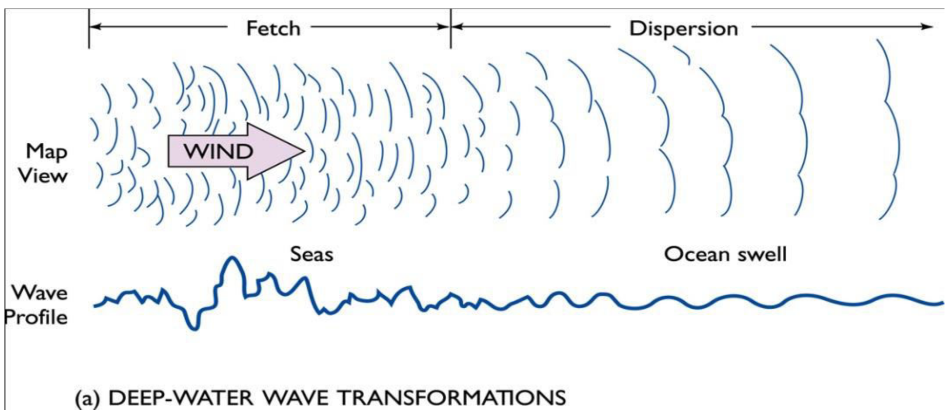


Fig.3: Transformation of sea in the generating area to swell.

7.4. Classification of waves:

Waves are classified according to their wave periods as shown in the table 1. The different forces that cause different types of waves and their restoring forces are indicated in Fig.4 which contains very smallest period "capillary" waves (with period less than 0.01 sec) to "ocean swells" (period about 15 sec) and tsunamis (period about 20 min). The amount of energy in a particular wave class is determined by the typical wave height in that class. The other higher order waves in this category are ultra gravity waves, gravity waves, infra-gravity waves, tidal and trans- tidal waves.

Table:.1: classification of waves basing on the period.

S.No.	Wave name	Period
1.	Capillary waves	< 0.01 sec
2.	Ultra gravity waves	0.01sec-1.0sec
3.	Gravity waves	1sec-30sec
4.	Infra Gravity waves	30sec-5 min
5.	Long waves	5 min-12 hrs
6.	Tidal waves	12 hrs-24 hrs
7.	Trans tidal waves	> 24 hrs.

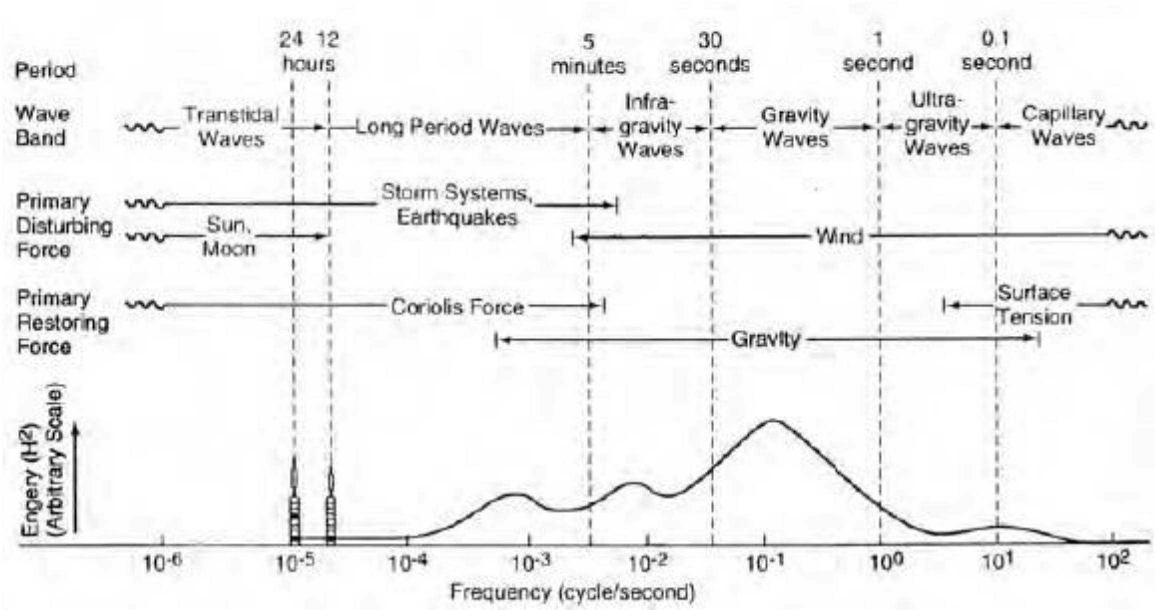


Fig.4. Classification of Wave spectrum basing on period (frequency)

7.5. Transformation of waves:

As the wave enters into the shallow water the transformations that occur are wave reflection, refraction and diffraction and then wave breaking. As the waves enter into shallow water, the wave length and velocity decreases and period remains same and wave height enormously increases as the wave energy is conserved and so the waves break at the shore.

7.6. Classification of waves basing on relative depth:

Ocean waves are generally classified according to a parameter called 'relative depth'. It is the ratio of the water depth (d) to the wave length (L). This ratio d/L is called the 'relative depth'.

The table 2 and Fig.5 gives Classifications made according to the magnitude of *d/L* and the resulting limiting values taken by the function *tanh (2πd/L)*. Note that as the argument of the hyperbolic tangent r = kd = 2πd/L gets large, the tanh (r) approaches 1, and for small values of kd, tanh (kd) tends to kd in the wave equation: $C^2 = \frac{g}{k}\tanh(kd)$ and $L = \frac{gT^2}{2\pi}\tanh(kd)$ ....(1)

in the case of shallow and deep water waves.

Table 2. Wave classification according to relative depth

Range of d/L	Range of r = Kd=2πd/L	Types of waves
0 to 1/20	0 to π/10	Shallow water waves (Long waves)
1/20 to ½	π/10 to π	Intermediate waves
½ to ∞	π to ∞	Deep water waves (short waves0

The simplifications which occur in the various wave equations arise by replacing the hyperbolic functions by their asymptotes for the particular range of relative depth are listed below in the table3.

Table 3. Approximations for hyperbolic functions

Function	Asymptotes	
	Shallow water	Deep water
Sinh kd	Kd	$\frac{e^{kd}}{2}$
Cosh kd	1	$\frac{e^{kd}}{2}$
Tanh kd	Kd	1

The deep and shallow water expressions can be obtained by using the simplifications resulted due to the asymptotes as given in table 3.

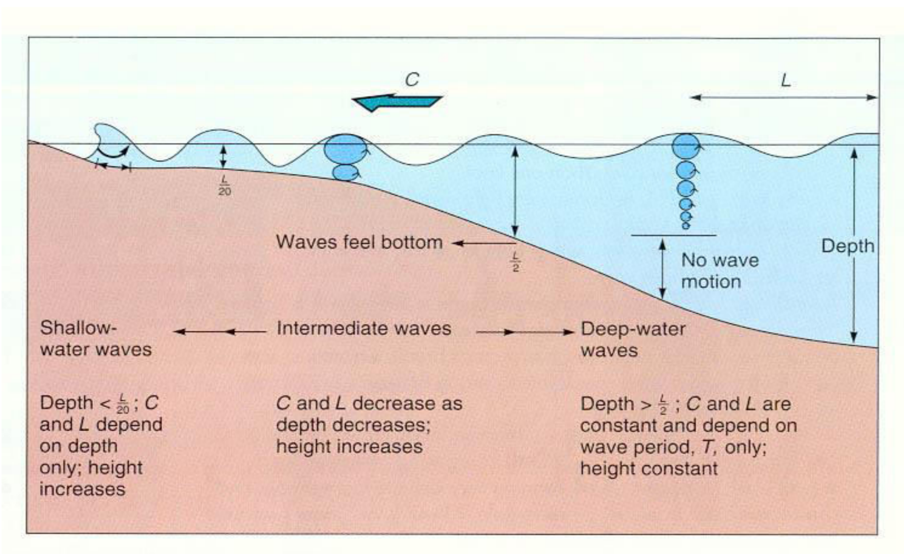


Fig.5. Classification of waves as they approach the coast from deep water and associated orbital motion of particles

7.6.1. Shallow water wave equation:

The shallow water condition from table 3 is Tanh kd → kd, so the wave equation

$$C^2 = \frac{g}{k}\tanh(kd) \qquad \text{……………………..(1)}$$

The wave equation 1 turns out in shallow water as:

C^2 = gd ………….(2)

This equation implies that wave velocity is independent of wave length and depends only on the water depth.

7.6.2. Deep water wave equation:

The deep water condition from table 3 is tanh kd → 1 and so equation 1, turns out to be

C_o^2 = g/k_o as we know k_o = 2 π/L_o; $$C_0 = \frac{gT}{2\pi} \qquad L_0 = \frac{gT^2}{2\pi}$$ ……….(3)

The suffix zero indicates the deep water conditions. The celerity and wave length for deep water conditions are therefore independent of water depth. One should note here that 'T' doesn't vary with the local depth of water.

If g = 9.8 m/s^2 and π = 3.14 then $\frac{g}{2\pi} = 1.56$

then C_0 = 1.56T and L_0 = 1.56 T^2 ………………..(4)

Question 1:

a) What is the speed of wave in deep water if its period is 20 seconds (Ans = 31.2m/s)

b) What is the speed of wave in deep water if its wave length is 312 m (Ans = 22.1 m/s)

c) At what speed will each of the waves referred in 'a' and 'b' travel in water of 12 meters deep.

Ans for 'c':

In the case of (a): T = 20 seconds, so L = 1.56 × 20^2 = 624 m

When d = 12 m, d/L = 12/624 = 1/52, according to table 4.1, this 1/52 comes under shallow water category ie.between 0 to 1/20. so shallow water wave velocity equation is to be used. Then the wave velocity according to equation 4.2 is $C = \sqrt{gd}$ = $\sqrt{9.8 \times 12}$ = 10.8 m/s

Similarly in case of (b), L = 312 m; so d/L = 12/312 = 1/26 which is also less than 1/20. so this also comes under shallow water category. So $C = \sqrt{gd}$ = $\sqrt{9.8 \times 12}$ = 10.8 m/s

This example implies that in shallow water where d/L < 1/20 all waves travel at the same speed when they reach to that particular depth.

7.6.3. Orbital motion of wave particles in a progressive wave:

In deep water, the orbital paths are circular in such a way that the orbits decrease exponentially as shown in Fig.6 since the relative depth d/L >1/2 and the radius of the orbit is ae^{kd}. However, the particle paths of shallow water waves are *elliptical* with a major semi-axis of A = a/kd and a minor semi-axis of B = a(d+z)/d The same condition is applicable to Intermediate waves as shown in Fig.7.

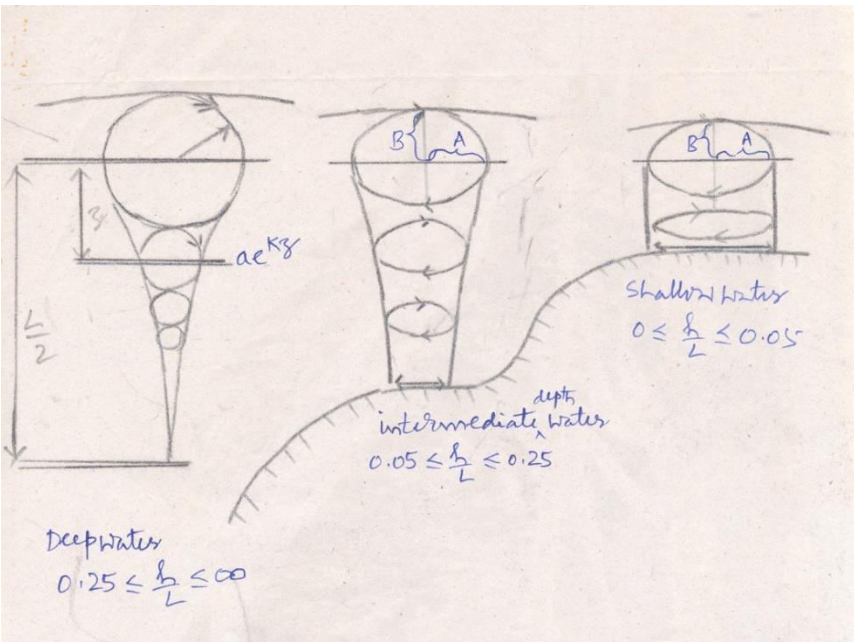


Fig.6 .Orbital motion of water particle in deep, intermediate and and shallow water waves (here Z = h = d is the water depth)

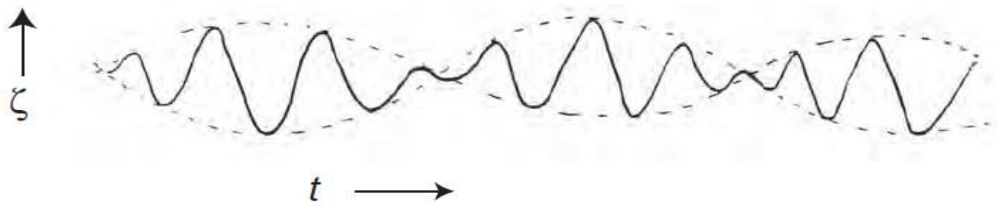
7.6.4. Differences between deep water waves (short waves) and Long waves (shallow water waves):

S.No	Character of the wave	Short waves (Deep water waves)	Long waves (shallow water waves)
1.	Velocity	Dependent on wave Length $C^2 = \frac{gL}{2\pi}$	Dependent on water depth $C^2 = gd$
2.	Movement of water Particles in a Vertical plane	Orbital motion of water particles is circular. The Radii of theses circles decrease with depth. Motion is imperceptible at a depth equal to half of its wave length as shown in Fig.6.	The motion is in ellipses. While the size of the ellipses decrease with depth and flattened, at a depth of half the wave length the ellipses become so flat like a line and so particles move back and forth as shown in Fig.6.

7.7. Group velocity:

If a wave profile is observed, it is not a single wave but it is a composite of many waves. That means in the ocean a mixture of many waves together travel. This mixture of wave profile is called a spectrum or

band as shown below. So a single wave has phase velocity (C) and the spectrum of waves has group velocity (C_g). This group velocity (C_g) in shallow water is half of the phase velocity (C) .



$$C_g = \tfrac{1}{2}C$$

while in deep water it is the ratio of the product of velocities of individual waves to the sum of it as:

$$C_g = \frac{C_1 \times C_2}{C_1 + C_2}$$

Where C_1 and C_2 are phase velocities of two individual waves.

From the above expression it is clear that the group velocity is half the phase velocity and $\frac{Cg}{C}$ can be seen to have the value ½ in deep water and approach to a limiting value of unity in shallow water.

As water becomes shallower, wave length becomes less important and wave speed depends only on water depth. As a result wave speed in shoaling water becomes closer to group velocity. Eventually at depths less than L/20, all waves approaching a particular depth travel at the same speed. So the group velocity and phase velocity at that depth are equal.

7.7.1. Wave Energy:

The energy possessed by a wave is in two forms. One is kinetic energy and the other is potential energy. Kinetic energy is the energy inherent in the orbital motion of the water particles and the potential energy is possessed by the particles when they are displaced from their mean position.

The total energy of a wave is partly potential and partly kinetic. The potential energy of a wave is equal to the work necessary to distort a horizontal sea surface into the wave profile which may be occurring beneath the sea surface which is given by the equation.

$$\text{P.E} = \frac{1}{4}\rho g a^2$$

Similarly Kinetic energy (K.E) of the wave equation is given as : $\frac{1}{4}\rho g a^2$

Total energy = P.E + K.E = $\frac{1}{4}\rho g a^2 + \frac{1}{4}\rho g a^2 = \frac{1}{8}\rho g H^2 \; (\because a = \frac{H}{2})$

7.7.2. ENERGY FLUX:

It is the rate at which the wave energy is transmitted in the direction of wave propagation. The rate at which the energy is propagated per unit length of the wave crest is called the wave power. So the energy flux has the unit of power.

$$\therefore P = E\, C_g\, N$$

where P is wave power, E is energy, C_g is group velocity and N is the shoaling constant. In deep water N = ½ and in shallow water N = 1.

Problem 1:

a) In deep water waves, what is the energy per square metre of a wave field made up of waves with an average amplitude of 1.3m ($\rho = 1.03\times10^3$ kgm^{-3})
b) What would be the wave power in KW/m of crest length if the waves had a steepness of 0.04 (1 watt = $1JS^{-1}$ and one kilowatt = 10^3 watts)
c) Also calculate wave power in shallow water.

Answer:

a) Energy $E = 1/8\ \rho g H^2 =$ $1/8 \times 9.8 \times 1.03\times10^3\times(2.6)^2 = 8.529\times10^3\ J/m^2$

b) Wave power: P = group velocity × wave energy per unit area

Given steepness = H/L = 0.04, we know H = 2.6; L = 2.6/0.04 = 65m

Phase velocity in deep water = C = $\sqrt{\frac{gL}{2\pi}} = \sqrt{1.56\times65} = 10.07 m/s$

Group velocity in deep water = half of phase velocity; $C_g = C/2 = 10.07/2 = 5.035$ m /s

Wave power = E.C.n = $EC_g = 8.529\times10^3\times5.035 = 42.7\times10^3$ J/m/s = 42.7×10^3 watts/m=42.7 Kw/m

c) wave power in shallow water = P = E.C.n = E.C as n = 1 in shallow water

= $8.529\times10^3\times10.07 = 85.4\times10^3$J/m/s = 85.4 kw/m

Problem 2:

A wave of period 10 seconds approaching the shore has a height of 1m in deep water. Calculate a) the wave speed and group speed in deep water b) wave length in deep water, c) the wave steepness in deep water d) wave energy, e) the wave power per meter of crest in water of 2.5 m depth.

Ans:

a) the deep water wave celerity, $C^2 = gL/2\pi$ or $C = gT/2\pi = 1.56\times T = 1.56\times10 = 15.6$ m/s
   Group velocity = $C_g = C/2 = 15.6/2 = 7.8$ m/s
b) $L = gT^2/2\pi = 1.56 \times 100 = 156$ m

c) Wave steepness = H/L = 1/156 = 0.0064

d) Energy $E = 1/8\ \rho g H^2 = 1/8 \times 9.8 \times 1.03 \times 10^3 \times (1)^2 = 1.26 \times 10^3\ J/m^2$

e) wave power in 12 m depth = we have to find whether deep water or shallow water
d/L = 2.5/156 = 0.016 which is less than 1/20 (0.05), so it is shallow water

So wave power in shallow water = P = E.C.n = E.C, as n = 1 in shallow water

$E.C.n = EC = 1.26 \times 10^3 \times 15.6 = 19.66 \times 10^3\ J/m/s = 19.66 \times 10^3$ watts/m=19.66 Kw/m

TIDES

Tides are the rhythmic rise and fall of the sea surface. They are caused due to gravitational attraction of sun and moon. The rising level is called flood tide and the falling level is called the ebb tide. There are two classes of tides basing on the period between flood and ebb tides. They are semidiurnal and diurnal tides. In the case of semidiurnal tides the duration of high tide (flood) to low tide (ebb) is 6 hours. That means two high tides and two low tides occur in the regions of semidiurnal category. When the duration between flood and ebb is 12 hours, they are called diurnal tides. This occurs due to sea basin topography and dimensions of the ocean.

7.8. TIDE PRODUCING FORCE:

The tide producing force (hatched thick arrow in Fig.7) is the resultant force between gravitational and centrifugal forces between earth and moon.

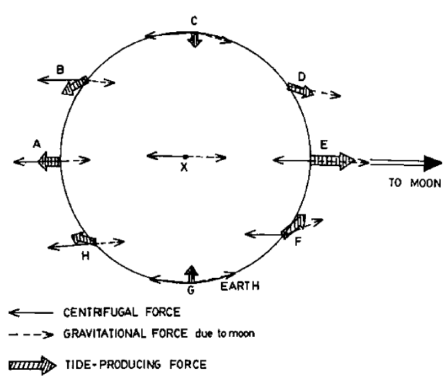


Fig.7 Tide producing force is the resultant of gravitational and centrifugal forces between earth and moon

From Figure 7 above it is clear that the magnitude of the gravitational force exerted by the moon on the earth is not the same at all points on earth, because all these points are not at the same distance from the moon. While the gravitational force of the moon is largest at point 'E', it is the weakest at point 'A' as Point E is closest and A is farthest from the moon. So while the gravitational force

is dominant in one half of the earth facing the moon, the centrifugal force is dominant than gravitational force in the other half of the earth away from the moon. In addition, the direction of the moon's gravitational force at all points on the earth is directed towards the center of the moon. So only on the equatorial diameter the gravitational force (dashed arrow in Fig.7) is straight and at other places it is inclined. As the tide producing force (hatched thick arrow in Fig.7) is the difference between gravitational and centrifugal forces, it also is directed in the same way as that of gravitational force.

Similarly Sun also acts in the same way on the earth's surface. But as the Sun is far away its influence is relatively small when compared to that of the moon.

The horizontal (F_H) and vertical components (F_v) of tide generating force are given as:

$$F_V = \frac{Gm\mu}{R^3} 3\left(\cos^2\varphi - \frac{1}{3}\right)$$

$$F_H = \frac{Gm\mu}{R^3}\frac{3}{2}\sin 2\varphi$$

These formulae show that the tide producing force of the moon at any point on the earth depends on the zenith angle of the moon (ϕ) and the moon's distance from the earth (R). As the R value at perigee (357,000 km) and apogee (407,000) is different, tidal forcing is not constant always.

The vertical component F_V attains its maximum value at $\phi = 0$ and 180^o when the moon is in zenith and Nadir points.

The horizontal component F_H attains its maximum value when $2\phi = 90^o$ or 270^o or $\phi = 45^0$ or 135^o.

The vertical component of tide producing force (F_V) is in the same direction of gravitation and therefore it doesn't have any effect on the water particle and so is neglected. Whereas the horizontal component of the tide producing force (F_H) will not be affected by gravity and so it will tend to drag the water particle in the ocean horizontally. This is called the tractive force.

Fig.8b shows how the horizontal component of the tide producing force is distributed over the earth's surface when the moon is on the equatorial plane. The water particles facing the moon are dragged towards it and the other particles (shaded area) are dragged away from it. On any latitudinal circle (A_1A_5) the variation of tractive force is significant. At point A_1 facing the moon it is directed south. When the point has moved to point A_4 the force is zero and when it has moved to A_5 the force is again equal and opposite. This explains why two high tides (A_1 and A_5) occur and two low tides (A_4 and its opposite) occur in one full rotation of the earth.

The sun produces a similar effect but relatively small tide producing force compared to the moon due to far off distance. When the two forces of sun and moon coincide it produces a maximum tide generating force and it is called spring tide. This happens twice during a period of 29.5 days. When these

two bodies act right angles the tide generating force is minimum. This minimum force produces neap tide. This also happens twice during 29.5 days.

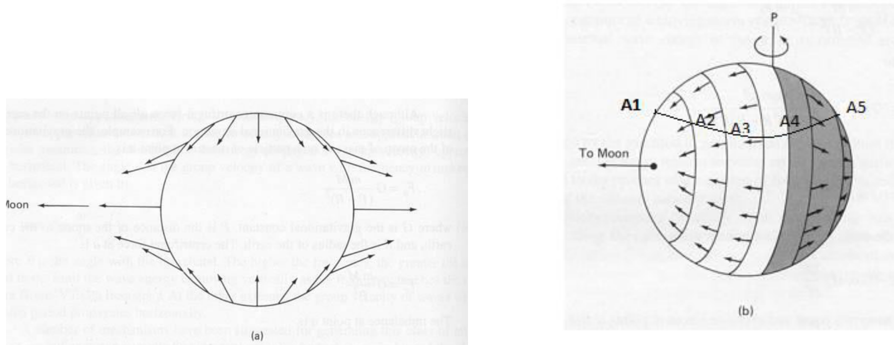


Fig.8.a) Distribution of the total tide generating forces for a cross section of the earth. b) the horizontal component of the tide generating force at the earth's surface

7.9. TIDE PREDICTION:

Tide predictions are based upon harmonic analysis of tidal data. The harmonic analysis of tides is based upon an assumption that the rise and fall of the tide at any locality can be expressed mathematically by the sum of a series of harmonic terms having relation with movement of the astronomical bodies. The general harmonic equation of the height of (H) tide can be written as :

$$H = H_0 + A\cos(at+\alpha) + B\cos(bt+\beta) + C\cos(ct+\gamma) + etc.$$

where H_0 is the highest of mean water level. Each cosine term in the equation is known as a tidal constituent. The coefficients A,B and C are the amplitudes of the constituents and are derived from observed or measured tidal data at each locality. The expression in parenthesis is called its phase angle which varies with time. The coefficient 't' (of a,b,c) represents the rate of change in the phase and is called the speed of the constituent and is expressed in degrees per hour. These constituents are independent of the place and are derived from astronomical data concerning the relative movements of the sun, moon and the earth. Table below gives some of these constituents.

Type of tide	Name of the tide	Symbol	Period	Theoretical amplitude (M_2 = 100)
Semi-diurnal Tides	Principal lunar tide	M_2	12.42	100
	Principal solar tide	S_2	12.00	46.6
	Lunar elliptical tide	N_2	12.66	19.1
	Luni solar declinational tide	K_2	11.97	12.7
Diurnal Tides	Luni solar diurnal	K_1	23.93	58.4
	Lunar declinational	O_1	25.82	41.5
	Solar declinational tide	P_1	24.07	19.3

The prediction of tide at a given locality by means of the above formula involves a good deal of calculation. In order to make the computation easier Lord Kelvin devised a machine in 1872 called tide predicting machine. By means of this machine the time of occurrence of high and low waters at a place can be calculated and printed in a book in every year and can be distributed to sea ports, navigators, sailors, shipping authorities and researchers. However, this gives only astronomical tide and the actual tide at a locality is little different because of the influence of basin topography and bottom stress which varies time to time and is not included. So the tide also should be measured with the help of a machine called tide-gauge. The difference between the observed tide and the predicted tide is called 'the Residual Tide'. If the residual tide is positive, it reflects on the storm surge as that is caused due to some terrestrial force like wind, barometric pressure or density.

7.10. TYPES OF TIDE:

There are different types of tides. Low tide, High tide, diurnal, semidiurnal and mixed type of tide, spring tide and neap tide etc.

With regard to the number of times of occurrences of tide they are classified as Diuranl, semidiurnal and mixed type. If one High tide and one Low tide occur in a day (24 hours) it is called diurnal tide and if two high and two low tides occur in a day then they are called semidiurnal tides (fig.9). If both are mixed they are called mixed tides as shown in Fig.10.

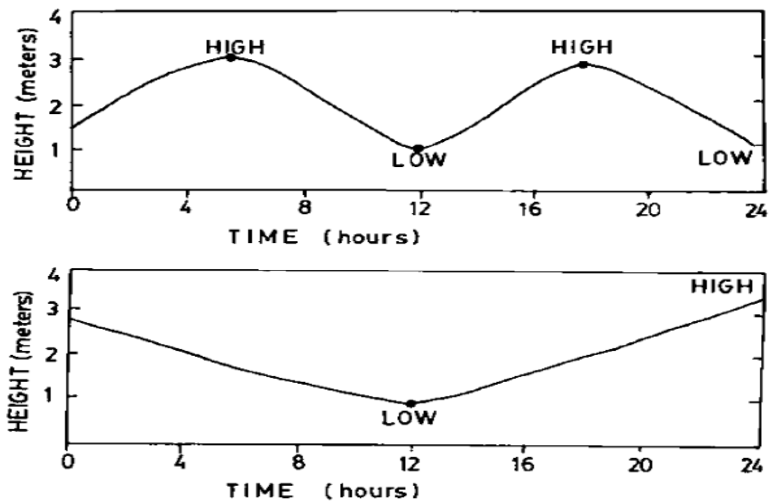


Fig.9. Shapes of semidiurnal and diurnal tidal curves in a day

For example,

i) Immingham (England), semidiurnal
ii) SanFrancisco (U.S.A), mixed type but dominant semidiurnal
iii) Manila (Phillippines), mixed type but dominant diurnal type.
iv) Do-San (Vietnam), full diurnal type. Here the period from one high tide to next high tide is 24 hours 50 minutes.

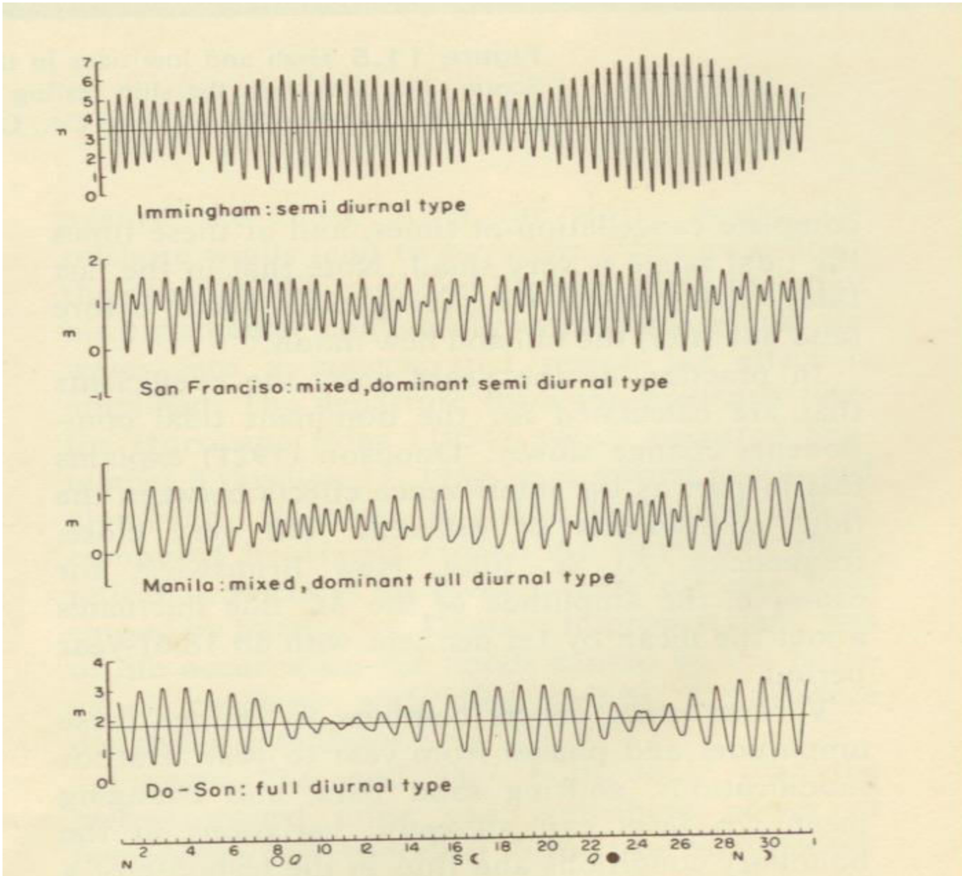


Fig.10.Example of shape of tidal curves in case of semidiurnal, diurnal and mixed tides at different places. The phases of the moon are indicated on the days in the x axis. ‘N’ shows the maximum northern declination of the moon; ‘S’ shows the maximum southern declination and ‘Q’ the time when the moon crosses the equator.

At any place, the tide generally has two cycles in a lunar month (29.5 days). In the lunar month, the earth faces full or new moon and the moon is in quadrants. When the moon is full or new, we get spring tides (highest tides) and when the moon is in quadrants, we get neap tides (Lowest tides) as shown in Fig.11.

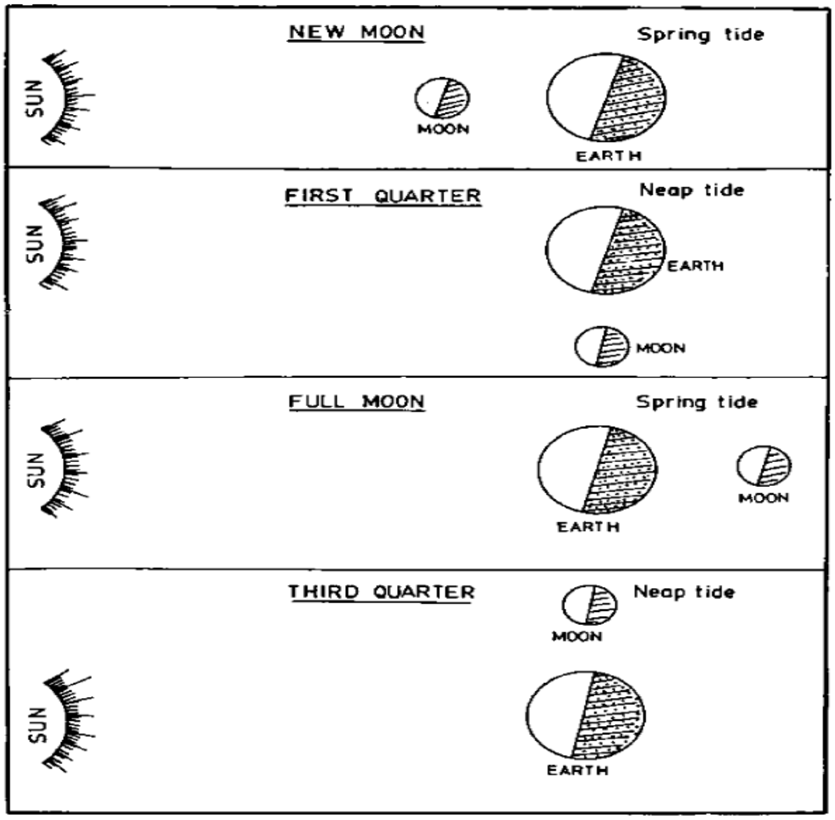


Fig 11. Spring and neap tides depending on the position of the moon

HANDOUT OF ARBAMINCH UNIVERSITY

INSTRUCTOR: Prof.(Dr.)A.S.N.MURTY

OCEANOGRAPHY & MARINE METEOROLOGY
MHD 0904
B.Sc. II Year: II Semester

Chapter – 8

Oceanographic Instruments and methods of observation

8.1. WATER SAMPLERS:

In order to find the water properties like salinity, density, PH, dissolved oxygen, nutrients (phosphates, nitrates, silicates) and other water quality parameters of sea water, these samplers are used to collect water samples at the surface and at different depths of the ocean. One of the most important water samplers is Nansen reversing water bottle.

8.1.1. NANSEN REVERSING WATER BOTTLE (NRWB):

It has a cylindrical barrel of capacity 1.2 liters. It is made up of brass and painted either white or yellow as shown in Fig.1 It has two plug valves on either side of the barrel and is connected by a connecting rod. The bottle is firmly fixed to the wire rope with a thumb screw clamp at the bottom while it is attached to the wire rope at the top with an automatic releasing mechanism. In the set position the tap will be above and air vent will be below. Once the messenger hits the head of the bottle, it is released and reversed. After reversal the two plug valves are automatically closed and so the water is trapped in the bottle in between the two plug valves.

First the bottle is lowered with the help of a rope in set position to the required depth as the bottle is in open condition in this position. After the bottle is reached to the desired depth, a messenger is released along the wire rope which will go and knock the bottle at its head where the releasing mechanism is there. As there is a firm grip at its bottom, when it is released at the top, it is reversed and the bottle is closed. Thus water sample at that depth is collected. Afterwards the bottle is hauled up and water is drained into glass bottles through the tap for analysis.

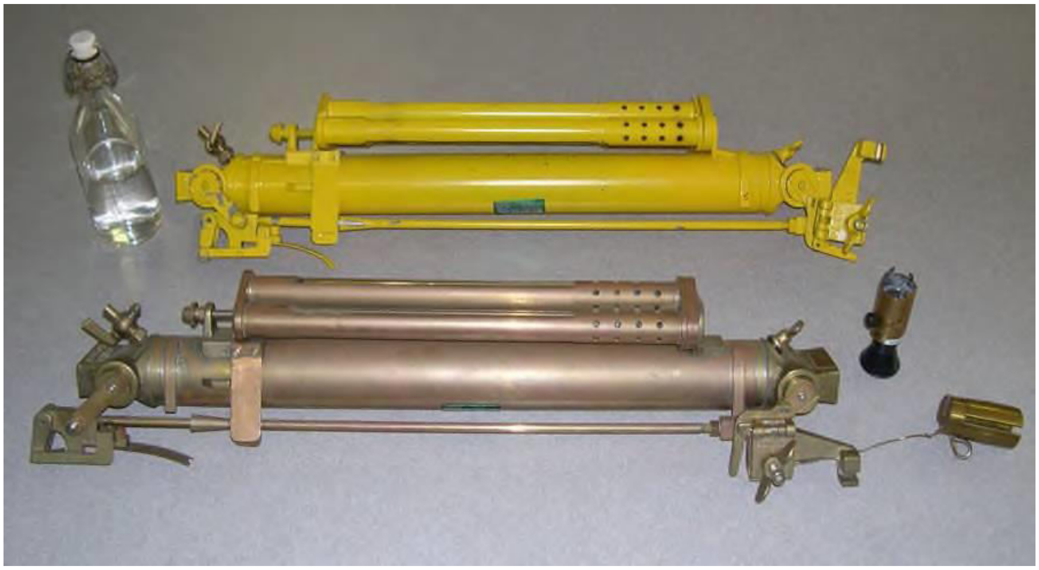


Fig.1. Nansen Reversing water sampler

8.1.2. NISKIN WATER SAMPLER:

The second type of water sampler is Niskin sampler as shown in Fig.2. This bottle is made up of PVC pipe and rubber covers on either side. The principle of working is same as that of Nansen Reversing water bottle. Only difference is as the Niskin sampler is made of PVC it is very light and easy to handle and even if it is lost it is less cost.

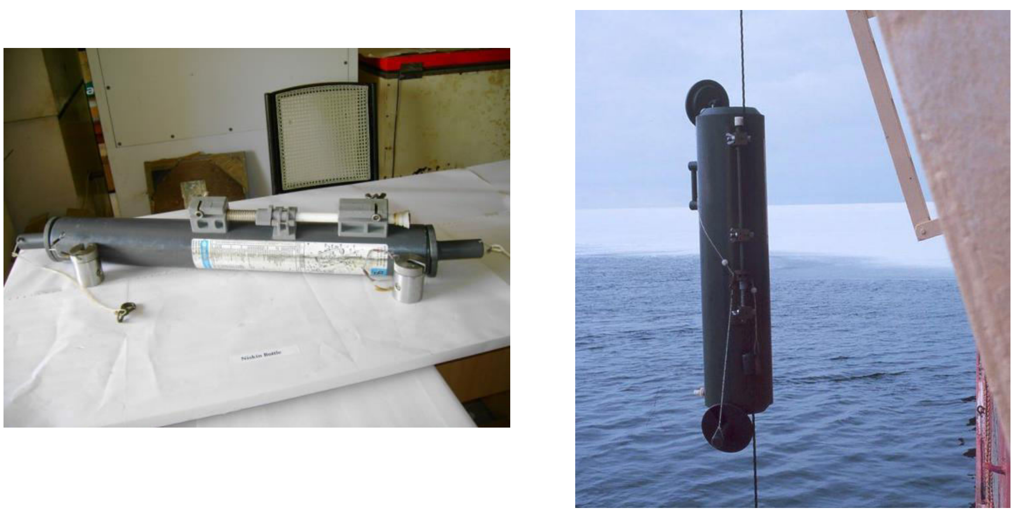


Fig.2. Niskin water sampler. Right side shows the position at the time of lowering in the sea from a ship

8.1.3. MESSENGERS:

These are the different types of weights called messengers (Fig.3) used for operation of sub surface instruments like Nansen reversing water bottle, MBT, Ekman current meter etc. There are three types hinged, slotted and open types

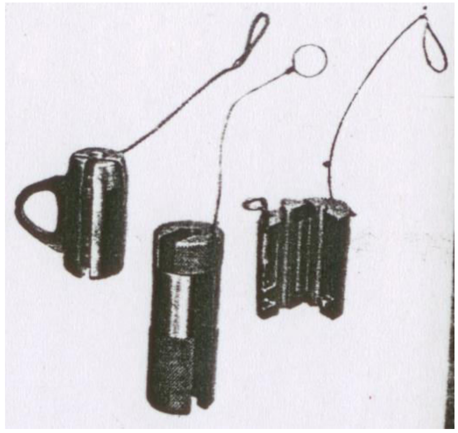


Fig.3. Messengers

8.1.4. PETERSON GRAB:

The grab looks like the one that is shown in the Figure 4. It has two jaws which can easily be opened and closed with the help of two cross cissored rods. The grab will be lowered with the help of a rope tied to these rods. This grab is used to collect sediment and sedentary animals from the sea or estuarine bottom.

Fig.4 .Peterson Grab

8.2. TEMPERATURE MEASURING DEVICES:

8.2.1. REVERSING THERMOMETERS:

There are two reversing thermometers as shown in Fig.5. One is Protected Reversing Thermometer (PRT) and the other is Unprotected Reversing Thermometer (UPRT). These two thermometers are attached to the Nansen Reversing Water Bottle. As these thermometers record temperature and depth of the water sample at the time of reversal, they are called reversing

thermometers. While the PRT measures the *in situ* temperature, UPRT measures depth at which PRT measured the temperature. So both the thermometers are simultaneously used and attached to Nansen Reversing water bottle..

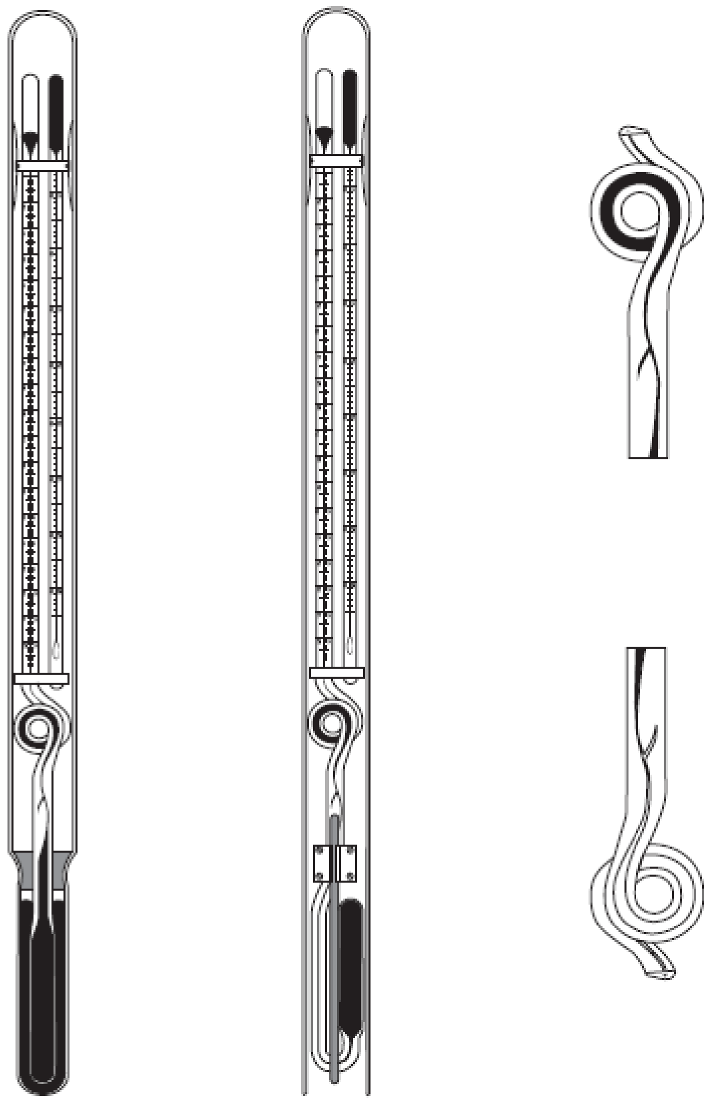


Fig.5. Protected (PRT) and unprotected (UPRT) Reversing thermometers. The spiraling and branching is in the third figure.

8.2.2. PROTECTED REVERSING THERMOMETER (PRT):

As shown in Fig.5, it contains a main reservoir through which the glass column runs through spiraling and branching with a graduated scale. Adjacent to this an auxiliary ordinary thermometer is fixed. Both these thermometers are enclosed with an outer glass jacket for protection from sea pressure. Once the PRT is sent to the required depth, it records the temperature of the water at that depth and on reversal it cuts off from the main reservoir and

extra mercury will not pass into the column due to spiraling and branching and so the meniscus is not disturbed while it is hauled up. After it is brought on to the deck when you see it in the set position, you can see the reading that it has recorded before the reversal.

8.2.3. UNPROTECTED REVERSING THERMOMETER (UPRT):

Its design is similar to that of protected reversing thermometer. But the only difference is the bulb is kept open without any protection by sea pressure. It has a main thermometer with looping and branching with a main reservoir at bottom and a secondary reservoir at the top in the set position. Adjacent to it an auxiliary thermometer is kept. Both the thermometers are enclosed in glass jacket which is open at the main reservoir. Because of the open end of UPRT when the thermometer is sent into the ocean at the desired depth, the value recorded by this thermometer is due to two effects. One is temperature effect and the other is sea pressure effect. As a result if we subtract the temperature effect we can get the effect of pressure. The sea pressure is nothing but the depth at which the sea water sample is collected. Both the protected and unprotected thermometers are enclosed to the Nansen reversing water bottle to know the temperature and depth of the water sample simultaneously.

By using both PRT and UPRT side by side we can get the *in situ* temperature at the depth it recorded.

By subtracting the value of PRT from UPRT, we get the depth of the sample at which it is collected.

8.2.4. BATHY THERMOGRAPH (BT):

There are two types of bathythermographs. One is Mechanical Bathythermograph (MBT) and the other one is Expendable Bathythermograph (X'BT).

MECHANICAL BATHY THERMOGRAPH (MBT): (Fig.6):

It looks like a torpedo with a nose in the front and fins at the rear end as shown in Fig.5. It consists of two elements. One is pressure element and the other is temperature element. Both are connected by a bourdon tube and stylus that marks on a smoked glass slide.

The pressure element consists of carpenter's bellows and the temperature element is a liquid in metal thermometer. The liquid used is xylene. Xylene is used here because it is sensitive for small changes of temperature and will not be solidified under subzero temperatures.

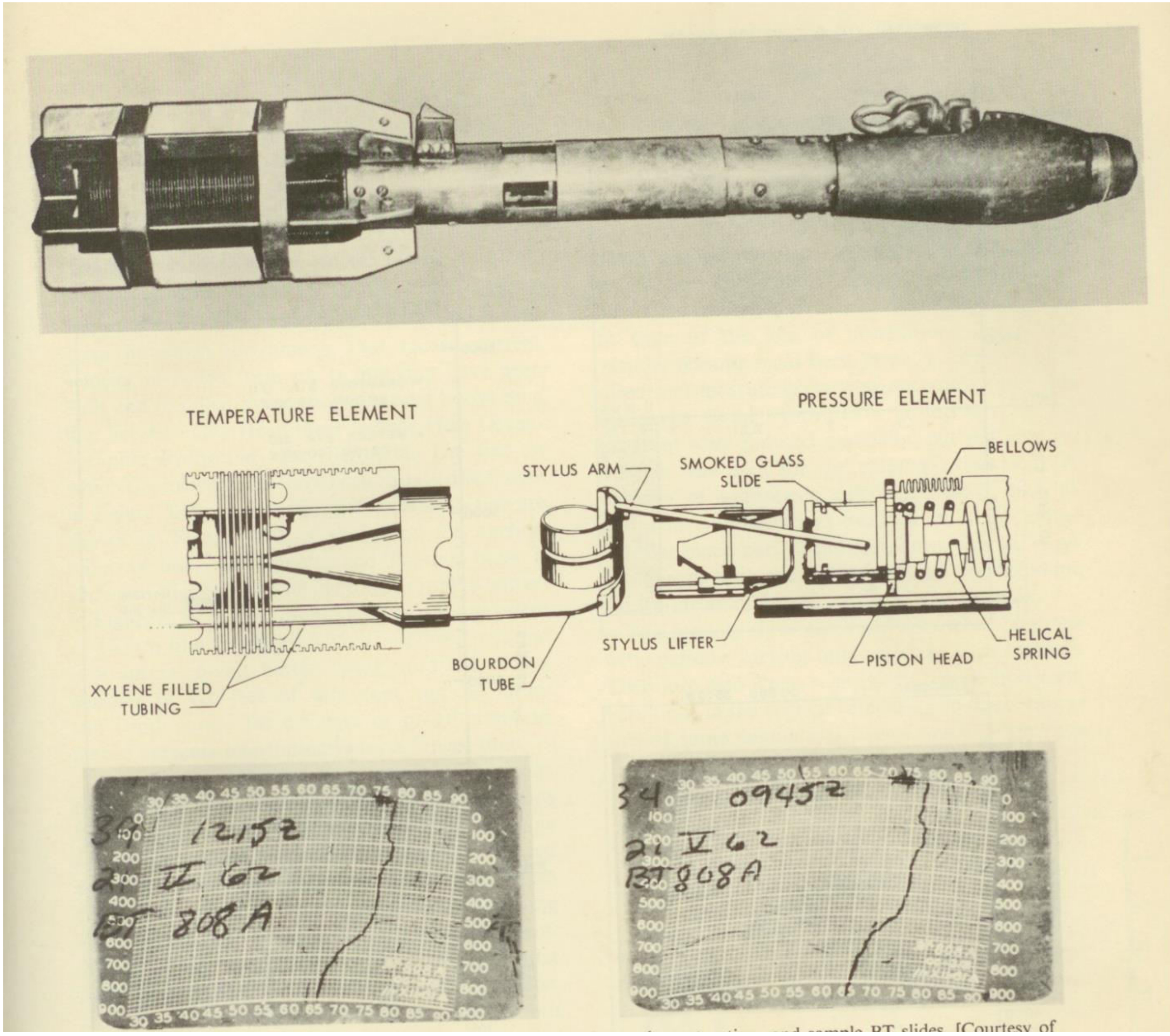


Fig.6. Mechanical Bathy thermograph. Middle: Temperature and pressure element. Bottom: the graph drawn by the metal stylus on the smoked glass slide.

At the center of both the elements on either side a small sachet is available to keep the smoked glass slide on which the stylus will mark the graph that represents the temperature versus pressure (depth). After the instrument is recovered on board, the smoked glass slide is removed from the sachet and is kept in a viewer and the graph can be seen on the screen of the viewer that can be manually read off.

DISADVANTAGE:

The serious disadvantage of this instrument is that when it is being hauled-up, there is a chance of erasing of the graph on the smoked glass slide due to entering of water. Then the entire exercise becomes futile. Another problem is there is a chance of losing of the instrument due to its heavy weight.

8.2.5. EXPENDABLE BATHY THERMOGRAPH (X'BT) (Fig.7):

To overcome the serious disadvantage of erasing out of the graph on the smoked glass slide in the MBT, X'BT is designed as shown in Fig.7. As the name implies it is expended (spent) once it is used. It contains a thermistor probe wound around with a lean copper wire of 500 to 1000 meters. This wire is connected to a deck computer which has a soft ware fitted to the respective probes. The probe is released into the ocean through a launcher that looks like a plastic gun. Once the probe is fired through the launcher, the probe enters the water and goes to deep layers as per the length of the wire (500 or 100 m). As it goes to different layers, the computer records temperature at different depths. Once the wire is finished, it is cut off from the computer and the probe is lost into the ocean. That is how it is expended. This instrument is preferred by everybody though it is expensive and lost every time of collection of data because we are sure to get the accurate data within a short span of time without much labor.

There was a world wide experiment in all the tropical oceans called Tropical Ocean Global Atmosphere (TOGA) program during 1980 – 85 to study the heat transport caused due to 1982/83 ElNino and other air-sea interaction problems. During this time X'BT has become popular and the data collected by this instrument simultaneously in all the three major oceans of Indian, Atlantic and Pacific has become invaluable.

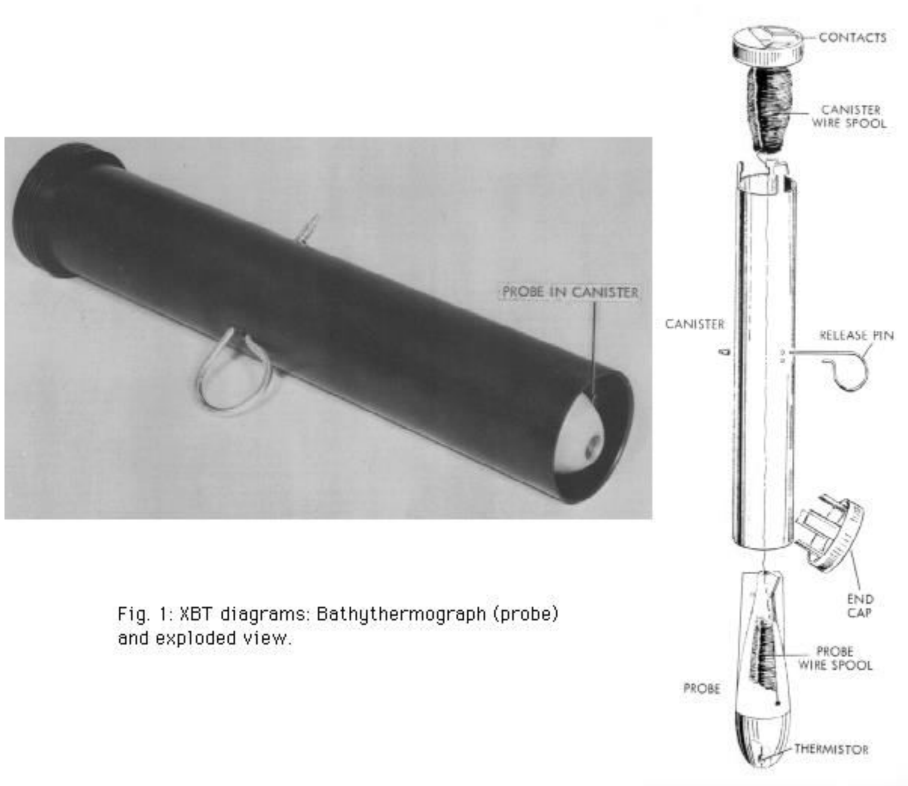


Fig.7A. Expendable Bathy thermograph probe (left) ; The releasing of probe from the container along with its canister tail wire that is connected to the computer.

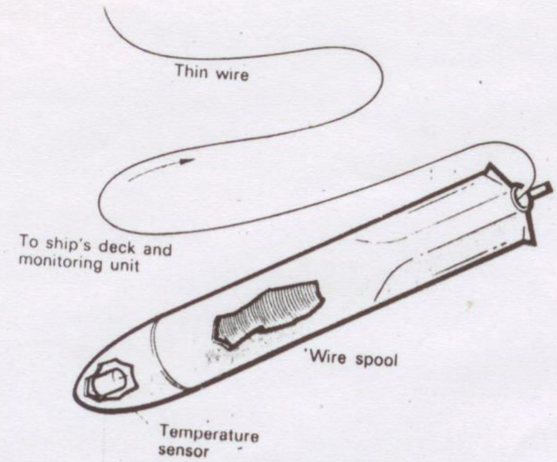
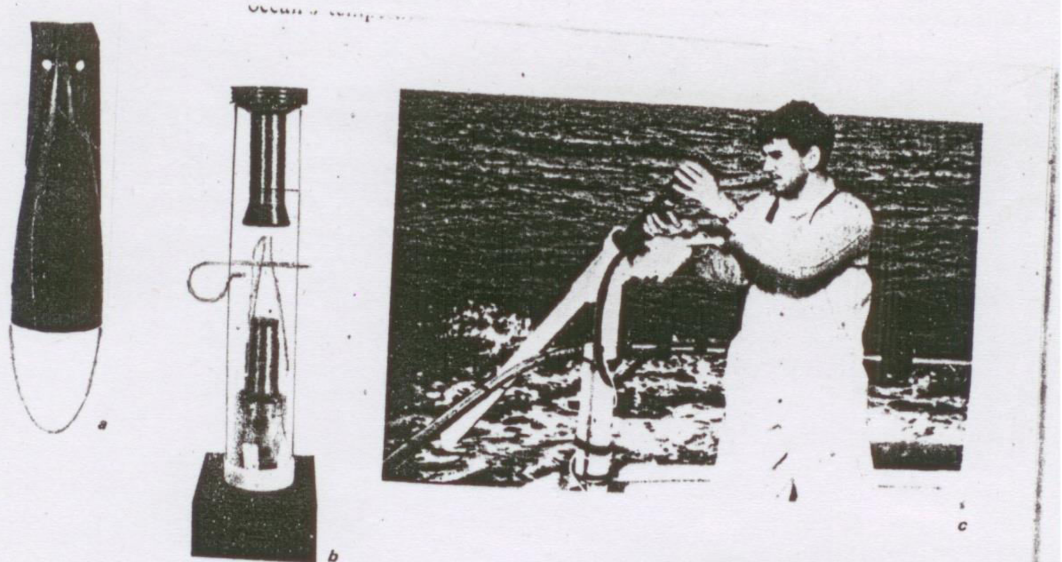


FIG. 4.8 *Diagrammatic representation of an expendable bathythermograph (XBT).*



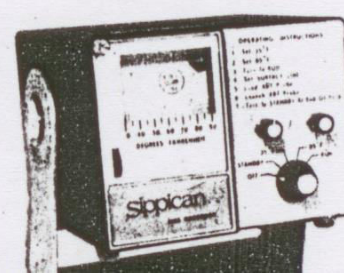


FIG. 4-22 *An expendable bathythermograph, or XBT. The key to this device is an expendable probe (a) that contains a thermistor or temperature-measuring device connected to a spool of fine wire; (b) a transparent view of (a). The probe is launched into the water (c) and wire is dereeled as the probe falls through the water. Changes in temperature are transmitted by the wire to a recorder (d) and since the rate of fall of the probe is known, depth can be read directly from the recorder. This device can be launched while the ship is moving or even from an airplane (in which case, a radio transmits the data to the plane). (Photographs courtesy of Sippican*

Fig.7B. X'BT probe, launcher from the deck of the ship and recording device

8.3. EXTINCTION AND TURBIDITY MEASUREMENT:

8.3.1. EXTINCTION COEFFICIENT:

The rate at which downward traveling light in the surface layer of the ocean diminishes is called extinction of radiation and is measured through a coefficient called extinction coefficient (K) which is computed with the help of the equation:

$$K = \frac{2.303}{L}\left(\log I_0 - \log I_z\right)$$

Where K is extinction coefficient, I_o is intensity of incident radiation, I_z is intensity of radiation at the required depth (z) and L is the distance between the two chambers of the hydrophotometer (usually fixed at one meter).

8.3.2. HYDROPHOTOMETER (FIG.8):

For measurement of extinction coefficient, the instrument used is called hydrophotometer as shown in Fig.8. It consists of two chambers. One chamber consists of source light at 230 V (I_o) and the other chamber consists of photocell that measures the received light (I_z). Both chambers are fitted with a carrier frame of one meter length. The instrument works with a car battery and the source and received intensities are measured with a deck instrument that contains a volt meter and a micro ammeter.

Using the above equation, extinction coefficient can be computed with the data of hydrophotometer.

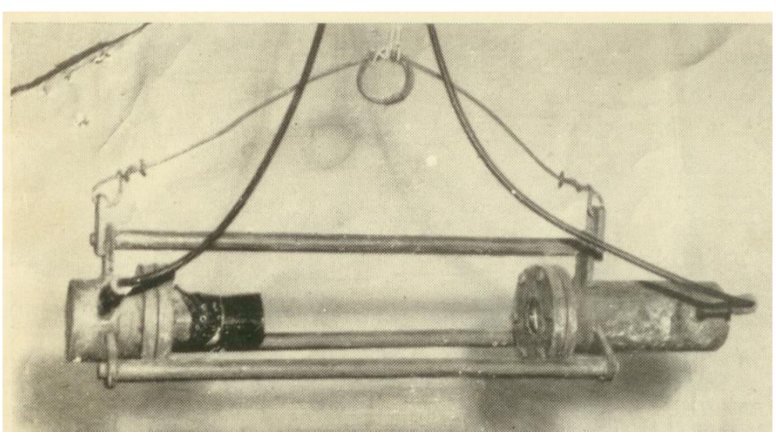


Fig.8. Hydro photometer

8.3.3. SACCHI DISC (FIG.9):

This is a brass disc of 30 cm diameter alternatively painted black and white for clear visibility as shown in Fig.9. It is lowered into waters of usually shallow water bodies like estuaries, lakes, lagoons and back waters. When this disc is lowered into waters vertically down with the help of a plastic wire rope tied with a lead weight, the depth of disappearance (d_1) and

reappearance (d_2) is measured. The average depth ($D = \frac{d_1 + d_2}{2}$) is called the Sacchi depth and the turbidity is calculated using the extinction coefficient $K = \frac{1.7}{D}$. Here K is directly related to turbidity. As extinction coefficient is more turbidity is more and vice versa. Using a calibration graph between K and turbidity (T), a regression equation can be established for a particular lake and regularly can be used for understanding turbidity.

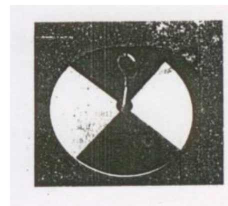


Fig.9. Saccchi disc

8.4. MEASUREMENT OF OCEAN CURRENTS:

8.4.1. EKMAN CURRENT METER (FIG.10):

It is an ingenious device prepared by a great oceanographer V.W.Ekman. It consists of a) propeller b) a compass and c) a vane.

The vane helps to orient the propeller against the direction of the water current when it is lowered to the required depth at which we want to measure currents.

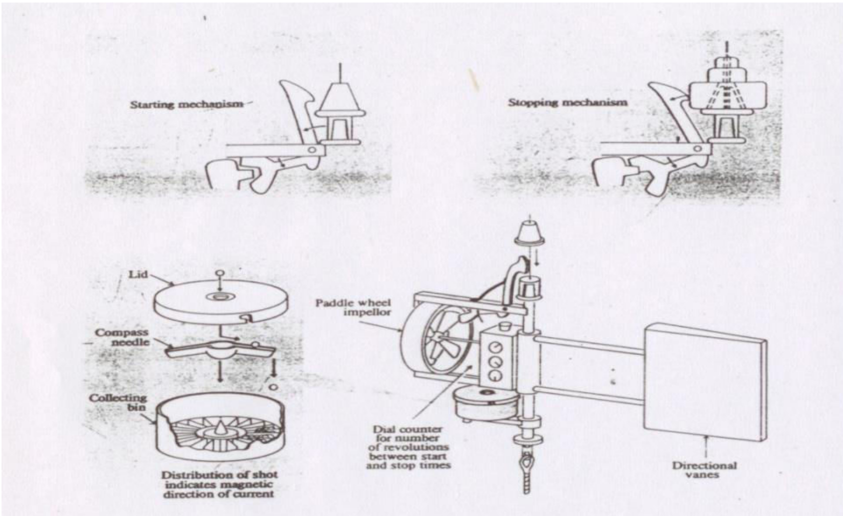
The propeller consists of a circular 10 cm diameter rim. The propeller is allowed to rotate at the desired depth by sending a metal weight called messenger along the wire rope which free the propeller from locking. A second messenger is dropped after few minutes to stop it. The number of revolutions made by the propeller is recorded by a dial box with a mechanical counter. The current speed is obviously proportional to the number of revolutions of the propeller per minute.

The current direction is recorded by the counting mechanism dropping lead balls (shots) at known intervals along a magnetic compass into a receptacle with 10^{o} sectors denoting eight directions. The numbers of lead shots that fall into each sector multiply by 10 and their addition will give the net direction of the current.

For example, the number of balls that fallen are as below:

Sector number	1	3	5	8
No.of balls	5	6	3	4

Then the resultant direction of the current is (5 + 6 + 3 + 4) $\times 10 = 180^{0}$.
Which means current is towards south.



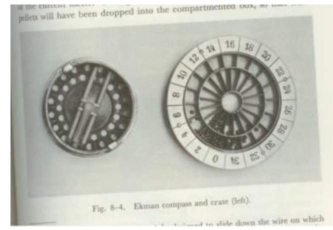


Fig.10A. Ekman current meter along with the receptacle and compass.

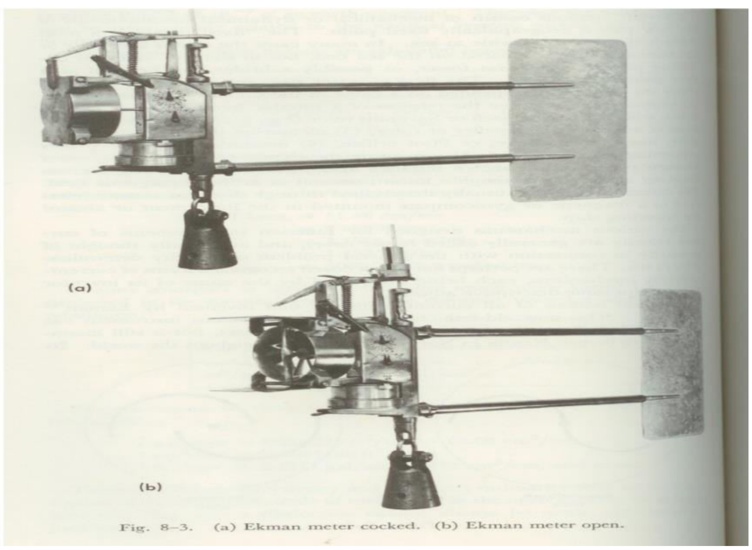


Fig.10B. Photo of Ekman current meter

8.4.2. ANDERRA CURRENT METER (Fig.11)

Now a days automatic recording electronic current meter which can be connected to computer is used as shown in Fig.11.

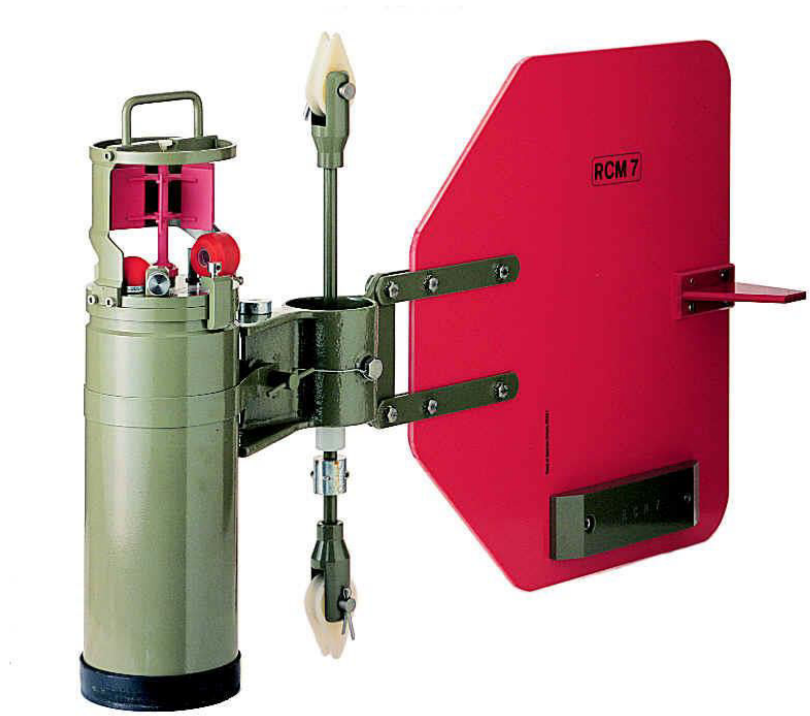


Fig.11. Anderra current meter

8.5. MESUREMENT OF TIDES:

8.5.1. TIDE GAUGE:

It consists of a float that is wound around a pulley with stylus on the clock drum that rotates along with the clock work. The float with the entire system rests in a well that is dug a few meters away from the sea coast. The variation of the tide (high tide and low tide) in the open ocean causes the variation of water level in the well. This variation of water level in the well is recorded on the graph paper wound on the clock drum. The graph paper is changed every day at a fixed time.

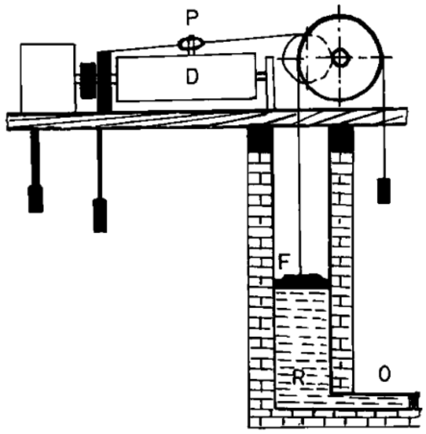


Fig.12. Tide gauge and its parts.

The tide gauge shown in Fig.12 consists of a float (F), which operates in the float well (R), to which the slow moving tide has free access. The more rapid waves caused by winds are filtered out by means of the small size pipe (canal) (O) connecting the well to the ocean. The rising and falling of the float (F) turns a wheel on the gauge which moves the pen (P) to and fro across a wide strip of paper on a drum (D) which is moved forward by clockwork. The combined motion of pen and paper gives a continuous graph showing the rise and fall of the tide. Generally this entire assembly will be housed in a shelter as shown in Fig.13. This shelter is built on a pier (a bridge on the ocean bank). The graph drawn by the tide gauge is monthly and so every month the chart is to be changed.

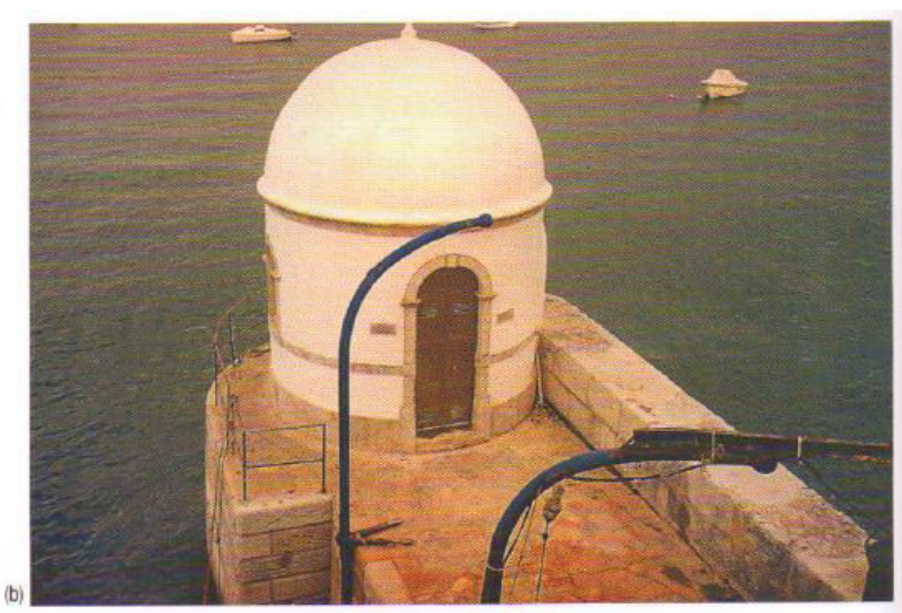


Fig.13. The pier (bridge) and the house where the tide-guage is kept in the back waters of the ocean.

HANDOUT OF ARBAMINCH UNIVERSITY

INSTRUCTOR: Prof.(Dr.)A.S.N.MURTY

OCEANOGRAPHY & MARINE METEOROLOGY
MHD 0904
B.Sc. II Year: II Semester

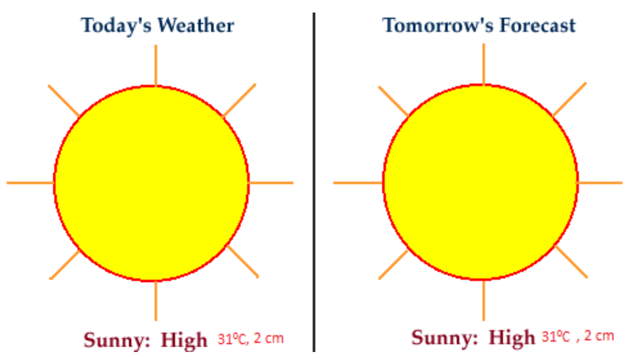
CHAPTER – IX
MARINE WEATHER ASSESSMENT & FORECAST

There are several different methods for marine weather forecasting. The method a forecaster chooses depends upon the experience of the forecaster, the amount of information available to the forecaster, the level of difficulty that the forecast situation presents, and the degree of accuracy or confidence needed in the forecast. The three different methods widely used are persistence method, trend method, climatological and Numerical Weather Prediction method.

9.1. FORECASTING METHODS:

9.1.1. Persistence Method:

Persistence method assumes that tomorrow's weather will be same as today. The persistence method assumes that the conditions at the time of the forecast will not change. For example, if it is sunny and temperature is 31^0C today in Arbaminch, the persistence method predicts that it will be sunny and 31^0C tomorrow also. If 2cm of rain fell today, the persistence method would predict two centimeters of rain tomorrow also assuming the same conditions.



The persistence method works well in mid latitudes where weather systems are rythmatic and go in sequential order. The weather patterns also without changing much move very slowly. It works very well in places like Ethiopia where the rainfall and thunderstorm come seasonally around the same time. However, if weather conditions change significantly from day to day, the persistence method fails to give a good forecast. The persistence method thus works very well only for shorter-term forecasts like one or two days in advance.

9.1.2. Trend Method

The trend method involves determining the speed and direction of movement for fronts, high and low pressure centers, and areas of clouds and precipitation using the previous history of the storm. Using this information, the forecaster can predict at some future time. For example, if a storm system is 1000 kilometers west of Ethiopia (Sudan) and moves to the east at 250 kilometers per day, using the trend method we can say it will arrive to Ethiopia in 4 days. The calculation is like this:

(1000 Kms/ 250 Kms per day = 4 days)

Using the trend method if a forecast is made only a few hours before, then it is known as "Nowcasting" and this method is frequently used to forecast precipitation. For example, if a line of thunderstorms is located 60 kilometers away northwest of Arbaminch and moving southeast at 30 kilometers per hour, we can say that the storms will arrive Arbaminch in 2 hours time.

Below is an example of using the trend method to forecast the movement of a cold front in U.S.A. Initially, the cold front moved 800 miles during the first 24 hours, from the central Plains to the Great Lakes (position 1 to 2 in Fig.2).

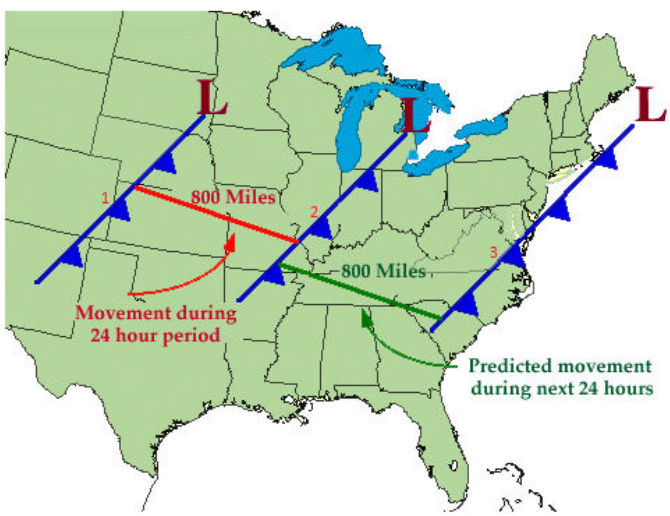


Fig.2. Prediction of the movement of cold front in United States of America

Using the trend method, we can say that this cold front weather system will move another 800 miles in the next 24 hours and reaches the East Coast of the United States (position 3in Fig.2). The trend method works well when systems continue to move at the same speed in the same direction for a long period of time. If they slow down, speed up, change intensity, or change direction, then the trend method of forecasting will fail.

9.1.3. Climatological method:

The Climatology Method is another simple way of prediction. This method involves averaging of weather data over many years. For example, using the climatology method we have to predict the temperature and precipitation of Arbaminch on July 4, 2016. Then we have to collect July 4th of

many years of the recorded data of temperature and precipitation from Arbaminch meteorological station and average them.

If these averages come out are 24^0 C with 0.18 cm of rain, then the weather forecast for Arbaminch on July 4th, using the climatology method, would be 24^0 C temperature with 0.18 cm of rain. The climatology method only works well when the weather pattern is similar to that expected for the chosen time of year. If the pattern is quite unusual for the given time of year, the climatology method will fail.

9.1.4. Numerical Weather Prediction Method:

Historical Back ground:

- The set of hydrodynamic equations which govern atmospheric motions has been known since 1858.
- But these equations could not be solved due to non linear parameters & closure problem.
- L.F.Richardson tried to solve by numerical means through finite difference scheme.
- But he encountered 2 problems- one is computational work is so great, the 2 nd one is small change in initial condition was amplifying to a large extent many times.

How Richardson's problems were solved now?

- Through judicious use of geostrophic approximation on over sensitive prognostic equations, first introduced by J.G.Charney.
- Use of super computer with main frame helped to solve the cumber-some calculation problem in short time.
- The closure problem was solved by parameterization concept of the mass, momentum and energy equations.

Numerical Weather Prediction (NWP) thus uses the computers and computer models to make a forecast. Complex computer programs, known as forecast models, run on supercomputers and provide predictions on many atmospheric variables such as temperature, pressure, wind, and rainfall etc.

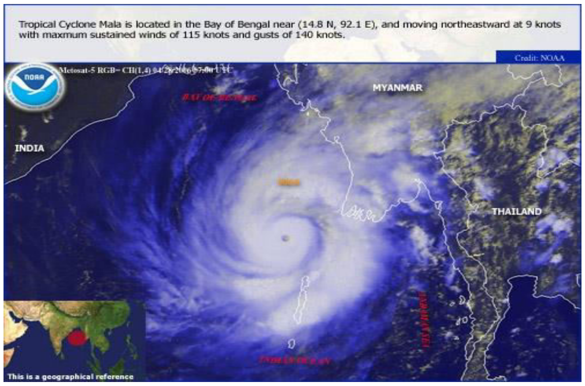
The defects in the NWP method are that the models used to simulate the atmosphere are not precise. This leads to some errors in the predictions. In addition, there are many gaps in the initial data as data from all parts are not available. If the initial state is not completely known, the computer's prediction of future state will be erroneous and not completely accurate. Despite these flaws, the NWP method is widely used all over world due to its convenience and availability of related soft-wares and computers everywhere with a hope to improve the errors in future.

Similarly tropical cyclones and the associated storm surges in coastal areas can be predicted using the above techniques.

9.2. CYCLONES AND STORM SURGE FORECASTING:

Tropical cyclones are low pressure centers that are developed due to excessive heating over tropical oceans. Usually they form in oceanic areas of temperatures more than 28^oc. It has 600 to

1000 km diameter and is about 10 km height. It has a centrally calm circular zone called ‘eye’ as shown in Fig.1. The wind speeds away from the eye will be of the order of 100 to 200 km/hr.



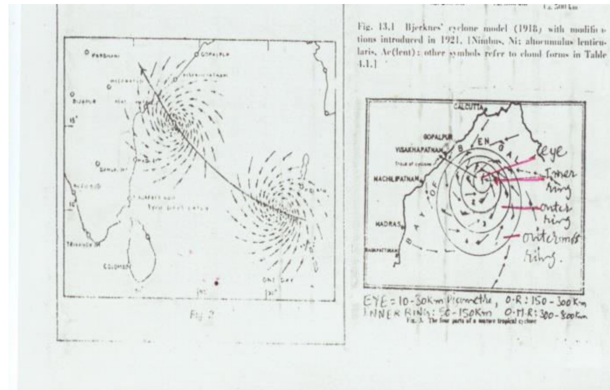


Fig.3. A mature tropical cyclone, the track and different rings of the cyclone

A mature cyclone gives torrential rains due to heavy cumulus and cumulonimbus clouds. At first it forms as a low in the deep ocean then moves towards the coastal zones away from the equator. During its passage it picks up latent heat in the ocean and develops into a cyclone or super cyclone depending on the conditions available.

The frequency of storms in the North Indian Ocean is from May to December with a peak in September and October. In the South West Indian Ocean region their frequency is during January to June and September to December. In the South Atlantic there are no storms and the South Pacific off Australia experience from December to March.

9.3. TROPICAL CYCLONE FOECASTING:

As tropical cyclones (T.C) are highly destructive because of high wind speeds and rain, it gives floods, storm surges and tidal bores in the river mouths and estuaries. So heavy loss of life and property takes place every year in the tropical countries bordering the oceans due to tropical cyclones. As they are incurring loss of several millions of dollars every year, tropical cyclone forecasting is essential. We must have four fundamental things while forecasting the Tropical Cyclones. i) the track, ii) the central pressure drop iii) eye character and iv) the intensity or size of the system.

Dvorak prepared a scale to understand the intensity of the cyclone basing on satellite cloud pictures. The wind speeds are derived from the cloud vector motions using Doppler radars.

Finally prediction of tropical cyclone implies prediction of its track. Because the track will exactly give the information of land falling (the point of crossing the coast), track prediction is important. This can be made basing on the above mentioned three methods of persistence, trend and NWP methods.

National Centre for Medium range Weather Forecasting (NCMRWF), New Delhi, India uses T-80 model for forecasting tropical cyclones using wind speed, cloud motion vectors, Sea Surface Temperature (SST), Central pressure drop, Radius of maximum wind etc.

9.4. STORM SURGES:

At the time of landfall (crossing the coast) of cyclone, a huge wave develops in shallow coastal areas and submerge the entire low lying area due to the impact of cyclone is called a storm surge as shown in Fig.4. Generally estuarine regions (meeting place of the river and ocean) experience this effect. This is known as storm surge. Generally the flooding into the villages and towns occurs due to two reasons. One is due to storm surge and the other is due to tidal bore. A tidal bore is the strong current developed against the flow of river due to moving cyclone in the ocean. Once tidal bore comes it swells on either side of the river banks. This kind of flooding causes lot of loss of life and property.

Heaps has defined storm surge as a raising or lowering of sea level caused due to cyclone.

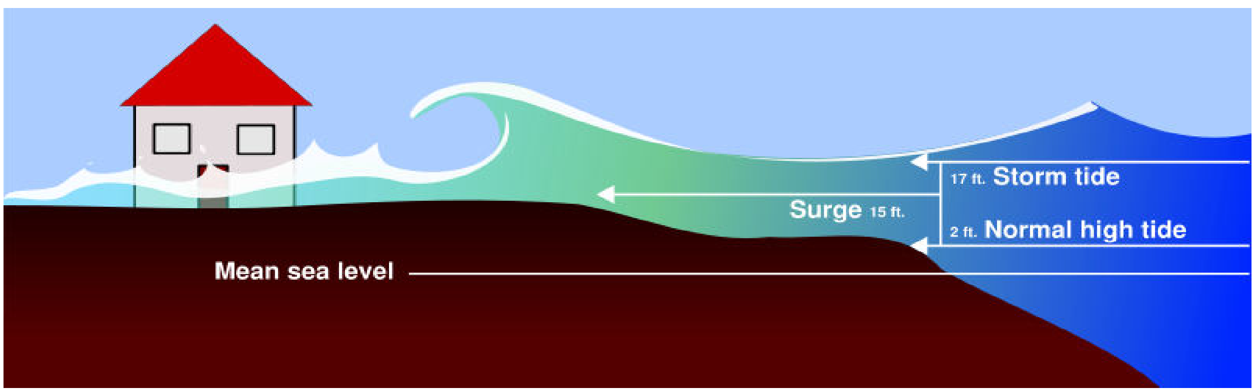


Fig.4. Advancing storm surge on to the coast

9.5. FORECASTING OF STORM SURGE:

The cyclone consists of barometric pressure drop and wind speed. Apart from this we have to take into consideration of steric sea level, tide and bathymetry also to forecast the accurate height of the storm surge as shown in Fig.5. Hence the equation is

$$H = B + W + S + T$$

Where H is height of the storm surge, B is the contribution due to barometric pressure drop, W is the wind surge (tractive force of the wind), S the contribution due to steric (density) level and T the contribution due to astronomical tide. If the storm surge comes at the time of high tide resonance takes place and so the surge will be enhanced.

We know the contribution of barometric forcing is that for every one millibar pressure drop one centimeter rise of sea level takes place. So by knowing the central pressure drop of the cyclone 'B' can be estimated. The sea level fluctuation due to density variation is known through seasonal distribution of physical properties that are usually available through oceanographic atlases or maps. Tide information is published by government organizations through tide tables and also measured by port trust and other organizations in every country. So only component that is not available is 'W'. The calculation of the contribution of wind is the main problem of calculation of storm surge.

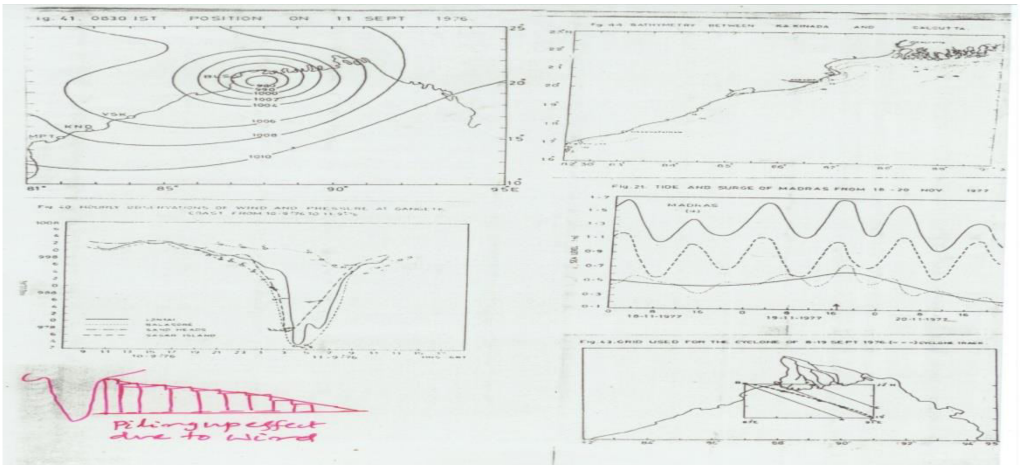


Fig.5 The factors contributing to enhance the storm surge are Radius, coast line curvature, inverted barometric effect, tide, surge pile up due to wind, surge along AB,CD,EF (bathymetry).

9.5.1. CALCULATION OF WIND SURGE DUE TO PILING-UP EFFECT:

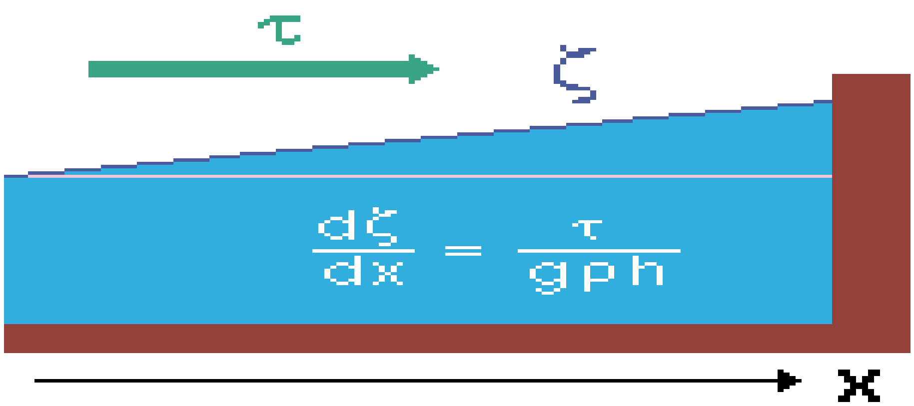


Fig.6. Calculation of wind set-up

The wind surge depends on wind set up (dz) which is given as

$$\frac{dz}{dn} = \frac{\lambda\tau}{g\rho_w h}$$

where dn is the orthogonal distance between any two isobaths h_1 and h_2 whose mean value is h. λ is the Mannings frictional factor varies between 1.0 to 1.5. τ is wind stress ($\rho C_d u^2$), ρ_w is the water density.

OCEANOGRAPHY & MARINE METEOROLOGY
MHD 0904
B.Sc. II Year: II Semester

CHAPTER – X
MARINE METEOROLOGICAL SERVICES – WMO

10.0. SERVICES FOR THE HIGH SEAS:

a) Provision of weather and sea bulletins
b) Marine meteorological support
c) Provision of information by radio- facsimile system
d) Marine climatological summaries scheme
e) Provision of special marine climatological information
f) Provision of marine meteorological information and expert advice.

To achieve the above objectives, the oceans and seas are divided into 8 areas for preparing the marine climatological summaries and with a view to continued international cooperation regarding the collection, archiving and exchange of marine data.

Members responsible for collection, archiving and dissemination of data are as follows:

Area	**Country**
North Atlantic	United Kingdom
South Atlantic	Germany
North Indian Ocean	India
South Indian Ocean & Antarctic & South East Pacific	Netherlands
North East Pacific	Japan
North & South West Pacific	USA

Apart from the above, two global collecting centers were given charge to collect data from the areas not covered and from fixed ships and also from global collecting centers. They are Germany and United Kingdom.

These responsible members make the data available to needy researchers and institution. The information regarding the availability of data etc is published and circulated by W.M.O time to time.

These members in collaboration with W.M.O prepare annual summaries, decadal summaries etc. and circulate on marine climatological information.

The facilities available for collection of data are fixed ship stations, moored buoys and fixed platforms in the deep sea areas. Apart from this many coastal stations also collect data on waves, tides, currents, SST, sea levels, sea ice, cyclones, storm surges, energy exchange etc.

Apart from this some world-wide experiments through GARP/FGGE like International Indian Ocean Experiment (IIOE), Indo Soviet Monsoon Experiment (ISMEX), Monsoon

Experiment (MONEX), Bay of Bengal Monsoon Experiment (BOBMEX), Tropical Ocean Global Atmosphere (TOGA) etc. for the Indian Ocean conducted. Similarly in Atlantic Ocean several experiments like METEOR, MODE etc are conducted to study North Atlantic hurricanes, mid oceanic ridge system, Gulf Stream rings etc. All these regional problems are studied as they have a global linkage. For example, the countries adjoining the Indian Ocean like India, Pakistan, Bangladesh, Sri Lanka, Myanmar, Indonesia, Maldives, and the East African countries are affected by the monsoons and tropical cyclones that are developed in the Indian Ocean area which is linked to ENSO. So to understand in detail the physical linkages, lot of data is to be collected and analyzed which is possible only through such global experiments in cooperation with the members of countries of W.M.O.

For distress and safety communication in the sea in 1973, the International Maritime Organization (IMO) adopted a recommendation called Safety NET scheme and all the countries follow this recommendation. Under this scheme shipping with navigational and meteorological warnings, meteorological forecasts, shore to ship alerts and other urgent information regarding the Safety Of Life At Sea (SOLAS) has been provided. Safety Net is a service through INMARSAT (International Marine Satellite). Safety Net messages can be originated and received by any registered user (eg. a member of WMO) anywhere in the world.

10.1. SERVICES FOR THE COASTAL & OFFSHORE AREAS:

a) Services under this category include coastal protection and coastal engineering works for gale, waves and storm surge warnings and also to assist the predictions of storm surges and coastal flooding.
b) Services for maritime search and resume operations
c) Services for special transport in coastal areas
d) Services for fishing
e) Services for fixed or floating installations and construction of harbors
f) Services for pollution monitoring and clean up operations.
g) Services for recreational facilities like boating, surfing, bathing etc.

Reference:

WMO No – OMM- 558